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Abstract

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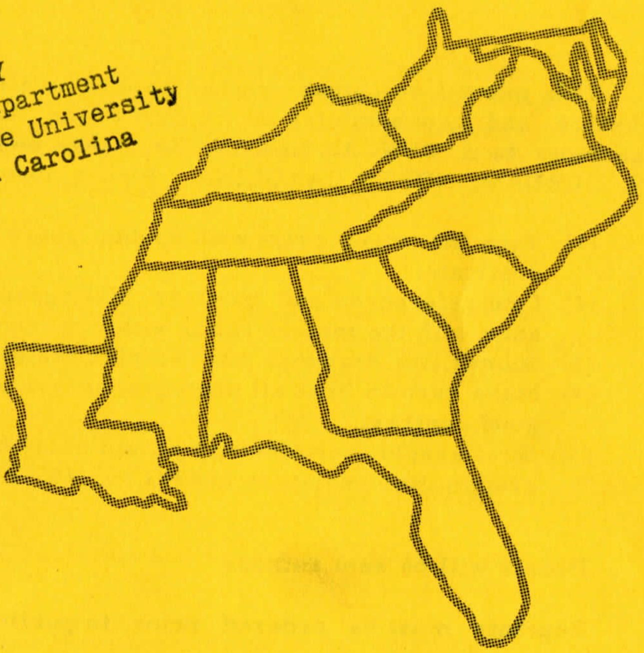

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SOUTHEASTERN GEOLOGY
Volume 21, No. 1
December, 1979

Memorial Issue
Dedicated to
George D. Swingle

Edited by Robert D. Hatcher, Jr.

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Figure 1. Dr. George D. Swingle and a group of students on a field trip near Knoxville, Tennessee, during the early 1960s.

GEORGE D. SWINGLE - IN APPRECIATION

George D. Swingle, as any teacher, affected the lives of every student he taught. To his doctoral students he was not only a teacher, but a companion and fellow collaborator in research. Because of our close association with Dr. Swingle while working under his supervision, each of us benefited considerably in our chosen professions as teachers, scientists, and indeed as persons.

George Swingle was one of the few teachers who could have a group of doctoral students waiting practically on the edges of their seats at the beginning of an advanced structural geology class. We would be waiting to hear some scenario ranging from topics like a dream about Appalachia always having been there (isn't it today?) to comparisons of windows in the Appalachians to those in the Alps. However, as each of us took his turn at presenting one or more topics, we wondered during their presentation if we had forgotten a fine point about which we would be reminded later.

Field trips with Dr. Swingle were enjoyed by all (Figure 1), from the rapid rate at which the distances between stops were covered to the many trips in rain or snow (an umbrella was considered standard equipment by some), to the amount of geology learned. It was chiefly in the field that we could partake of his bountiful knowledge of the geology of the Cumberland Plateau, Valley and Ridge and Blue Ridge, as well as some of his philosophy concerning the conduct of research.

As a research director George Swingle has the quality of knowing how much direction a student needed and would spend more or less time with him as the circumstances required. That quality was most beneficial.

The most important thing given us by George Swingle was a philosophy of how to do and understand geology in the field. It is a difficult thing to describe, but appears to surface in the modern concept of integrated structural-petrologic-stratigraphic-geochronologic studies of the core complexes of orogenic belts. His philosophy embodies the tool of detailed geologic mapping as the basis for data gathering and subsequent interpretation. Dr. Swingle firmly believed in this aspect but emphasized that a fine line must be drawn between spending too much time in one place and covering an area in a reasonable amount of time and with adequate data collection. In his slow mountain drawl and with tongue in cheek, he liked to draw the analogy of his two classes of geologists: "piddlers and leg men," to illustrate this principle.

We feel that today in many graduate geology programs in the U. S., field geology, even using modern techniques and principles, is being deemphasized or abandoned with the rationalization that field studies do not offer sufficiently challenging problems for advanced degrees. We believe the philosophy taught us by Dr. Swingle is somewhat unique and therefore offer these papers by his former doctoral students in an effort to honor him in a small way.

STRUCTURAL GEOLOGY ALONG THE EASTERN CUMBERLAND ESCARPMENT, TENNESSEE -- A SECOND LOOK

By

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ABSTRACT

Structure along the Allegheny front in Tennessee may be explained as variations of the thin-skinned model expressed by the Pine Mountain block. Near the structural front anticlinal structures are formed where the décollement in Cambrian shale rises along major tectonic ramps that cut diagonally across a thick carbonate section, and then flatten into the bedding of higher shales of Silurian and Devonian or Mississippian and Pennsylvaniana age. Structural cross sections, controlled by key oil tests and available seismic lines, suggest that strata potentially favorable for hydrocarbon production lie at shallow depth adjacent to and below the westernmost thrust sheet of the Valley and Ridge.

INTRODUCTION

The geology of the western part of the Valley and Ridge, along the Allegheny structural front in Tennessee, is known largely through the efforts of Professor George D. Swingle and his students at the University of Tennessee (Figure 1). Swingle was primarily interested in the areal distribution of rock units, the structure and strat graphy, and the economic geology of the region. Much of his work and that of his students along the structural front was published by the Tennessee Division of Geology in several geologic quadrangle maps (Swingle, 1960a, 1960b, 1960c, 1964a, 1964b, 1964c, 1964d, 1964e; Luther and Swingle, 1963; Swingle and Luther, 1964; Tiedemann and others, 1965; Milici and Swingle, 1972) and was of primary importance in establishing the cooperative geologic mapping program of the Division of Geology and the Geologic Services Branch of the Tennessee Valley Authority.

The area is of continued interest because of its potential for oil and natural gas, particularly from Mississippian formations in the Cumberland Plateau and in the adjacent Valley and Ridge. Clearly the geologic mapping begun by Professor Swingle over 20 years ago will

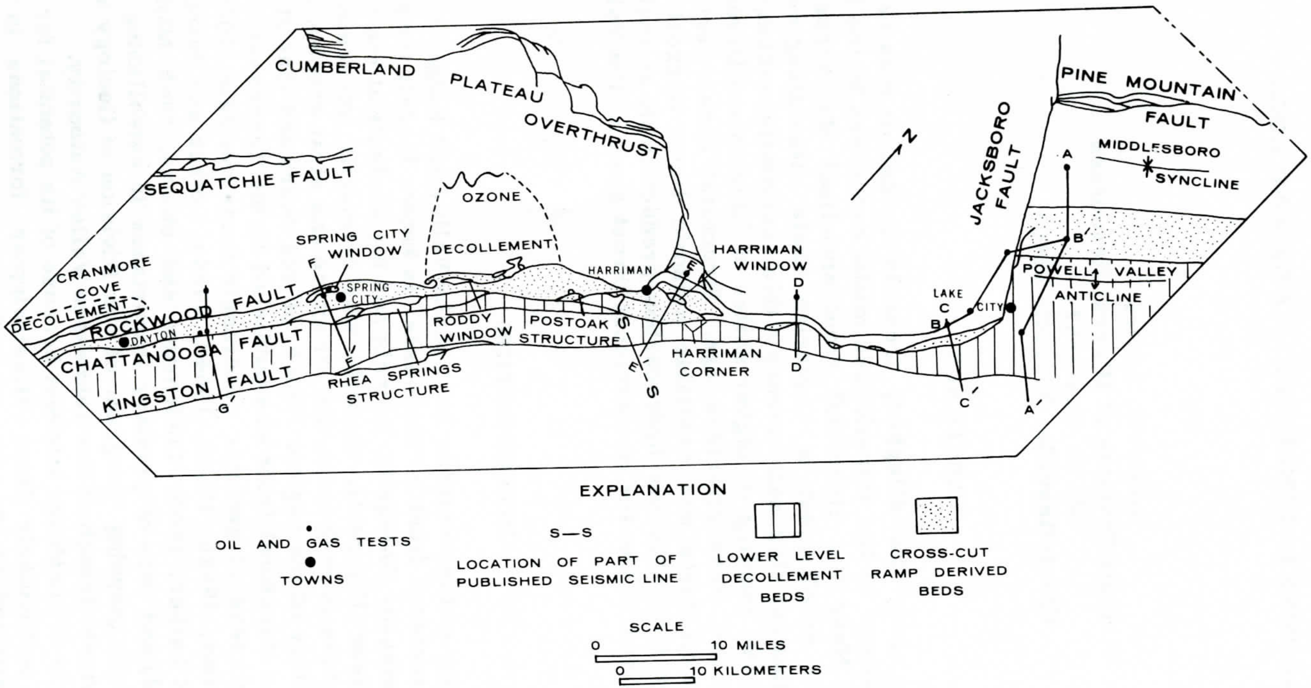

Figure 1. Generalized geologic map of part of the Allegheny structural front in Tennessee. The front follows the northwest edge of ramp-derived beds along the trace of the Rockwood fault and across the Pine Mountain block. LC - Lake City, H - Harriman, SC - Spring City, D - Dayton.

provide the framework for future oil and gas exploration in the region.

I first became involved in the Allegheny front project in 1954, while mapping the Harriman Corner for a Master's thesis at the University of Tennessee (Milici, 1962). Subsequently, I helped Swingle edit and prepare his geologic quadrangle maps for publication and helped him prepare illustrations for his summary discussion of the geology along the front (Swingle, 1961).

Swingle based his concepts of the tectonics of the region on analyses of the surface geologic relations shown by detailed geologic mapping. The purpose of this paper is to combine both surface and subsurface information, especially seismic data, in order to expand and amplify Swingle's interpretation of the folds and faults along the western part of the Valley and Ridge and in the adjacent Cumberland Plateau in Tennessee.

Acknowledgments

Reviewed by Wallace de Witt, Jr., and Leonard D. Harris. Research sponsored by Branch of Oil and Gas Resources, U. S. Geological Survey, Department of the Interior under U. S. G. S. Grant No. 14-08-0001-G-256.

RELATIONS OF STRATIGRAPHY TO STRUCTURE

To a considerable degree the structure of the western part of the Tennessee Valley and Ridge is determined by contrasting modes of failure of carbonate and terrigenous clastic lithologies within the stratigraphic sequence. In that region strata can be combined into five lithotectonic units[1] (Figure 2). A lower terrigeneous clastic unit (1) composed of Cambrian shales of the Rome Formation and Conasauga Group is the site for basal décollement beneath the Valley and Ridge. The overlying carbonate unit (2), composed of the Knox Group and of Middle and Upper Ordovician limestones, is cross-cut by major tectonic ramps. Overlying terrigenous clastic units of the Rockwood Formation and Chattanooga Shale (3), and of the Pennington Formation and Gizzard Group (5) are sites for upper level décollements, whereas the intervening carbonate unit of Mississippian limestones (4) is cross-cut by comparatively minor tectonic ramps.

Deformation of the lithotectonic sequence has resulted in an interrelated arrangement of subhorizontal décollements within the terrigenous clastic, shaly lithotectonic units, which are connected by tectonic ramps across intervening carbonate units (Figure 3).

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[1] Lithotectonic units are groupings of strata that were deformed together and in a similar way under the influences of the forces of deformation.

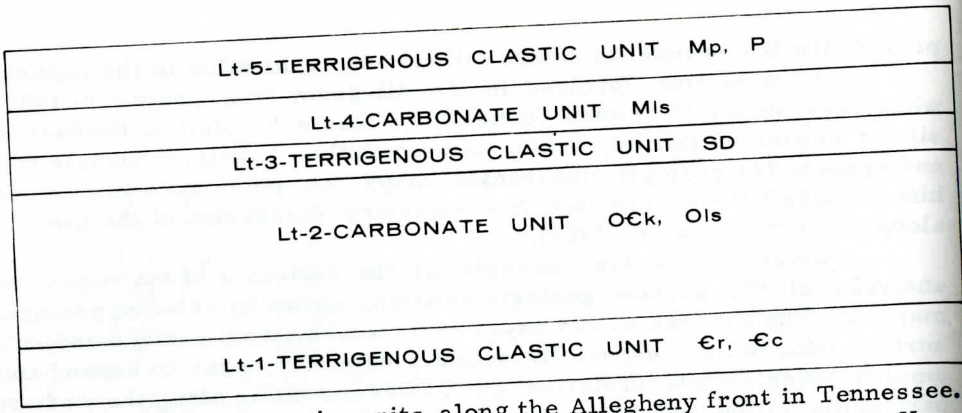

Figure 2. Lithotectonic units along the Allegheny front in Tennessee. Ꞓr-Rome Formation, Ꞓc -Conasauga Group, OꞒk- Knox Group, Ols-Middle and Upper Ordovician limestones, SD-Silurian (Rockwood) and Devonian (Chattanooga) shales, Mls-Mississippian limestones, Mp-Pennington Formation, P-Pennsylvanian formations.

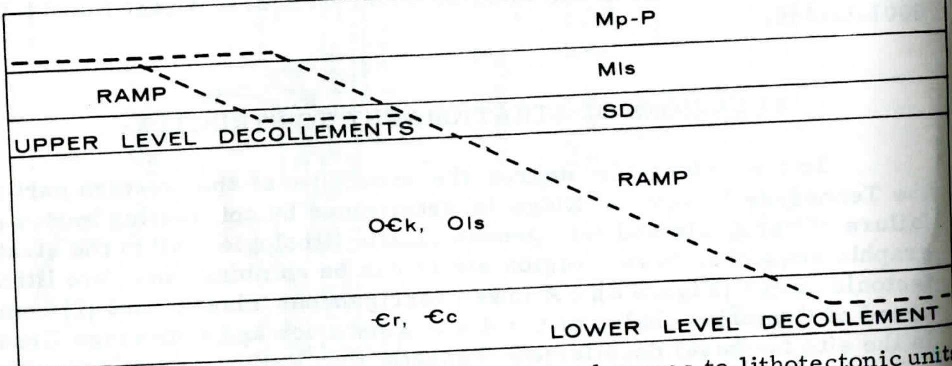

Figure 3. Relationship of décollements and ramps to lithotectonic units

THE PINE MOUNTAIN BLOCK--A MODEL FOR STRUCTURE ALONG THE ALLEGHENY FRONT

The structure of the Pine Mountain block was determined largely through the efforts of the U. S. Geological Survey (see Harris, 1970, for a summary). In brief, the Pine Mountain block consists of a broad shallow syncline of Pennsylvanian strata on the northwest (Middlesboro syncline), a broad flat-topped anticline of Cambrian strata on the southeast (Powell Valley anticline), separated by steeply-dipping, vertical or overturned beds of intermediate age (Figure 4A, 4B). This anticline-syncline couplet was formed by movement and duplication of beds above

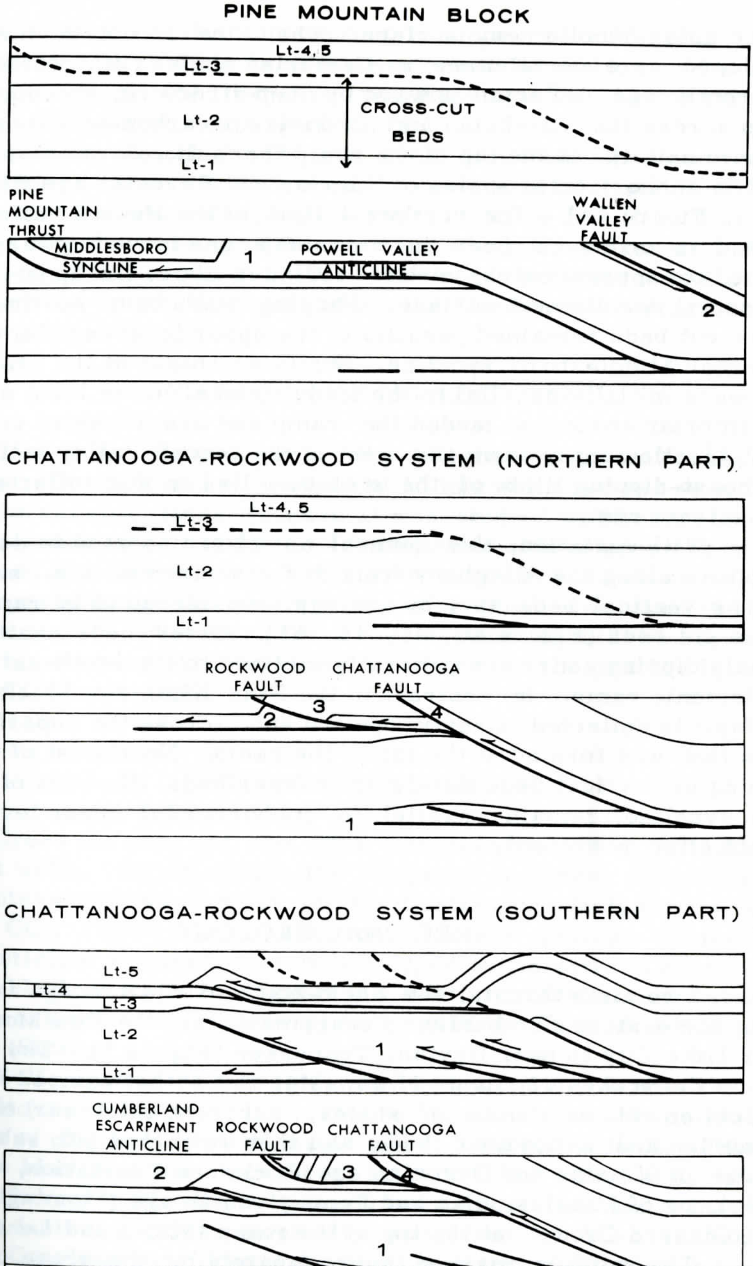

Figure 4. Comparative development of the Pine Mountain block and northern and southern parts of the Chattanooga-Rockwood Fault system. Lithotectonic units (Lt-1, Lt-2, etc. are explained in Figure 3; sequence of faulting is shown by numbers 1, 2, etc.

an irregular décollement surface. The Pine Mountain thrust, which developed as a décollement in Cambrian shales (lithotectonic unit 1), apparently was deflected upward by deep structure, thereby forming a ramp across the Cambrian and Ordovician carbonate sequence (lithotectonic unit 2). At the top of the ramp the fault relocated as a décollement in subhorizontal shales of Silurian and Devonian age (lithotectonic unit 3, Figure 4A). The northwest limb of the Powell Valley anticline formed as cross-cut beds in the ramp zone moved upward and were rotated to approximately vertical attitudes above the upper level subhorizontal décollement surface. Hanging wall beds northwest of the cross-cut beds remained parallel to the upper level décollement to form the broad Middlesboro syncline. Strata southeast of the crosscut beds that were initially parallel to the subhorizontal lower level décollement in Cambrian shales ascended the ramp and are repeated on the upper level décollement to form the crest of the Powell Valley anticline. The southeast-dipping limb of the structure lies on and reflects the dip of the tectonic ramp.

With variation, this general model can be used to decipher the structure along the Allegheny front to the southwest, i. e., steeply-dipping or vertical beds are, by analogy, considered to be ramp-derived cross-cut beds (Figure 4C, 4D, 4E, 4F). Older beds southeast of the steeply dipping zones are generally inclined to the southeast parallel to the tectonic ramp. In contrast to the Pine Mountain block, however, the fault is deflected steeply upward and across the superficial anticline that was formed at the top of the ramp. Northwest of the steeply dipping or vertical beds flat-lying younger beds, like beds of the Middlesboro syncline, remain parallel to subhorizontal upper level décollements after movement.

REGIONAL GEOLOGY

Two main thrusts, the Rockwood and Chattanooga faults extend along the eastern Cumberland Escarpment from the Pine Mountain block near Lake City to near Dayton, Tennessee (Figure 1). The faults constitute the broken remnants of a master thrust that ramped upward from a décollement in Cambrian shales, across a thick carbonate unit of Cambrian and Ordovician beds and then refracted into subhorizontal shales of Silurian and Devonian age (Rockwood Formation, Chattanooga Shale), or of Mississippian and Pennsylvanian age (Pennington Formation-Gizzard Group) at the top of the ramp (Milici and Leamon, 1975).

The ramp is marked approximately by the trace of the Chattanooga fault, which brings beds of Cambrian age to the surface along most of its length (Figure 1). The Rockwood thrust block lies northwest of the Chattanooga, and consists of steeply-dipping, vertical or overturned beds of Ordovician to Mississippian age which have been removed from the ramp position and transported westward a short distance along a nearly subhorizontal, upper level thrust. In places the thrust lies in shaly beds of the Rockwood Formation or Chattanooga Shale. Elsewhere the thrust follows Pennington shales or terrigenous clastics and coals of Pennsylvanian age.

With the exceptions of the Pine Mountain block and the Ozone and Cranmore Cove décollements, upper level thrusts northwest of ramp-derived strata extend only a short distance into the Cumberland Plateau along the bedding of Silurian and Devonian shales or in Mississippian and Pennsylvanian shales. However, movement was sufficient to produce the anticlinal trends along the Cumberland Escarpment.

GEOLOGIC CROSS SECTIONS

The cross sections across the Allegheny front in Tennessee presented herein are based on a study of 13 key wells and two seismic lines. One of the seismic lines (Figure 1) was studied by Harris (1976), and is the basis for interpreting the southeastern side of cross section EE'. The second line lies several miles southwest of cross section GG', and my interpretation of that seismic cross section was used in the construction of GG'. However, the cross section should be regarded as schematic, showing structural style rather than the precise location of folds and faults at depth.

PINE MOUNTAIN BLOCK

Cross sections of the Pine Mountain block are controlled by 6 different wells. Significantly, the Kingston fault near its northern end can be interpreted as a splay off the Chattanooga fault (cross sections AA' and CC', Figure 5). In addition, surface geology of the Powell Valley anticline suggests that it is warped sufficiently to accommodate a buried block in its core, thinner but similar to the buried Bales thrust block described by Harris (1970) for part of the Pine Mountain block in Virginia (Section AA').

Wells studied on a roughly north-south line across the Jacksboro fault (cross section BB', Figure 5) show virtually no change in elevation of footwall formations on either side of the fault. This confirms the idea that the Jacksboro fault does not offset beds below the Silurian-Devonian décollement level, but rather turns and joins the Pine Mountain fault at depth,as originally suggested by Rich (1934). The Sewanee Conglomerate (Pennsylvanian) is structurally high in the Lindsey Coal Company well on the southwest side of Jacksboro fault, but the Chattanooga Shale is not, suggesting that some splay thrusting accompanies the sharp upturning of beds on the southwest side of the fault, resulting in a duplication of beds, probably in the Pennington-Gizzard interval.

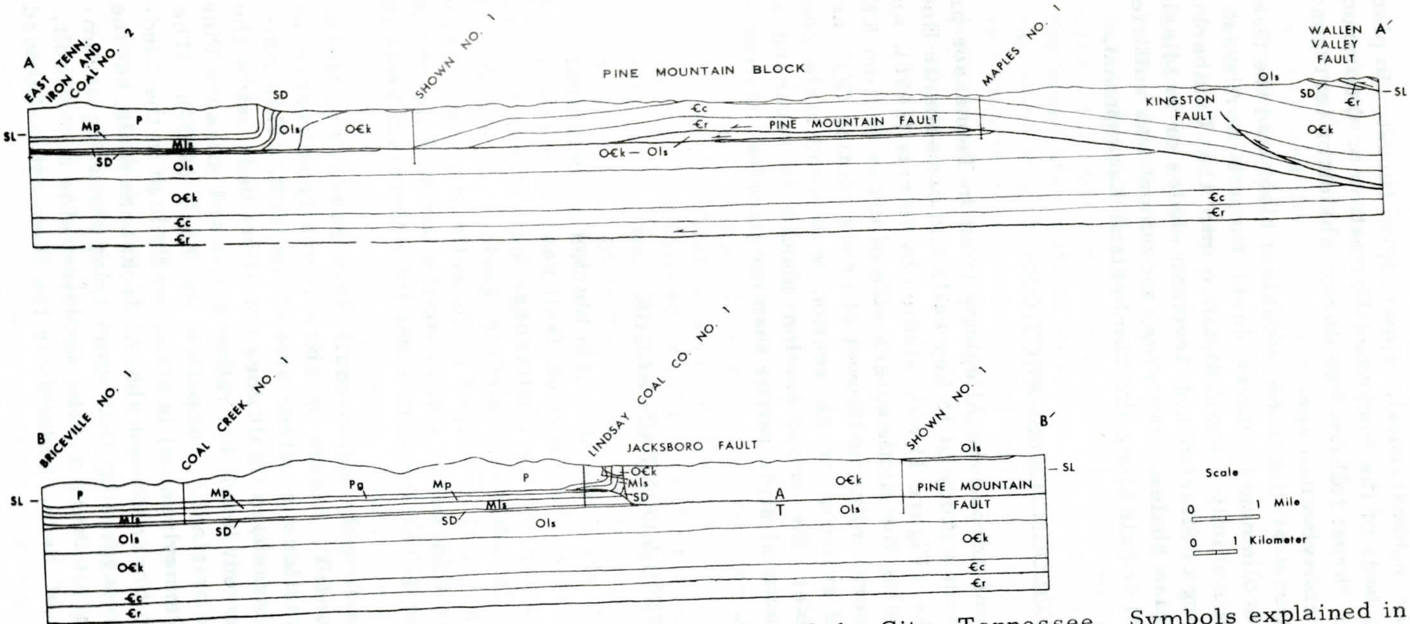

Figure 5. Cross sections of the Pine Mountain block near Lake City, Tennessee. Symbols explained in Figure 2.

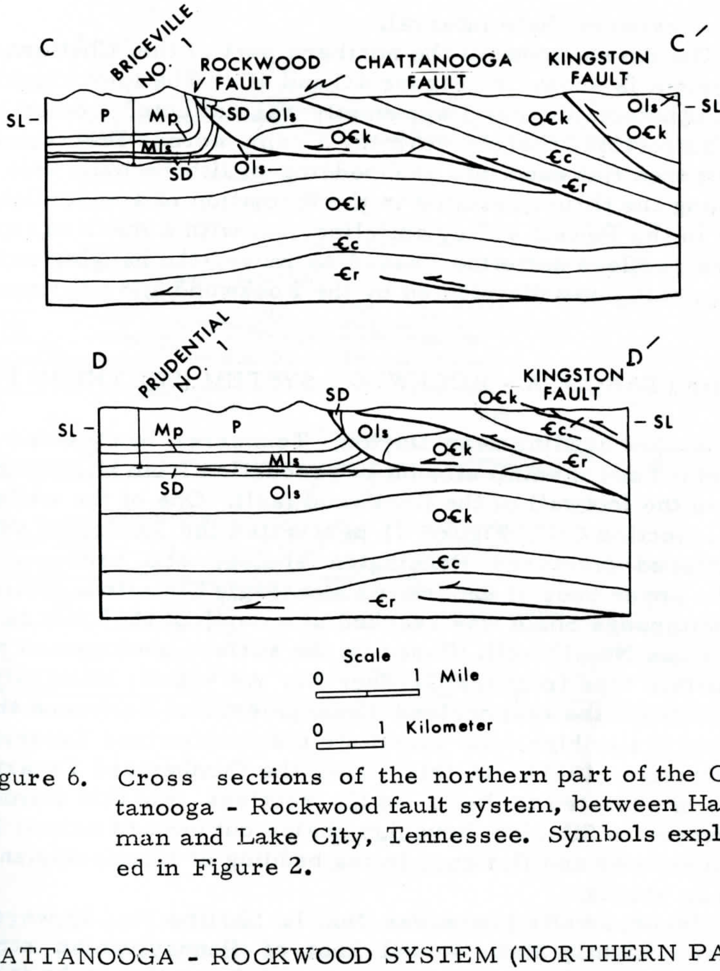

Figure 6. Cross sections of the northern part of the Chattanooga - Rockwood fault system, between Harriman and Lake City, Tennessee. Symbols explained in Figure 2.

CHATTANOOGA - ROCKWOOD SYSTEM (NORTHERN PART)

Cross sections CC' and DD' (Figure 6) are variations of the cross sections of the Pine Mountain block. The major difference is that the Pine Mountain block moved as a unit some 10-12 miles to the northwest (Harris, 1970, p. 166), whereas the block capped by Pennsylvanian beds southwest of the Jacksboro cross fault moved only a little above the décollement at the Silurian-Devonian level. This caused the Chattanooga and Kingston faults to cut towards the surface southwest of the Jacksboro fault in order to distribute the remainder of the 10 to 12 miles of displacement. The eastern edge of the Cumberland Plateau between Lake City and Harriman apparently has moved a little to the northwest. The Briceville No. 1 well, only a few thousand feet northwest of the Cumberland Escarpment (cross section CC', Figure 6) contains over 300 feet of deformed Chattanooga Shale (over 7 times the normal stratigraphic thickness), indicating that thrusting there is in the

Silurian-Devonian shale interval.

The development of the northern part of the Chattanooga-Rockwood system is shown in Figure 4C and 4D. The lower level décollement in lithotectonic unit 1 apparently was deflected upward by an anticlinal obstruction, thereby forming a ramp across lithotectonic unit 2. The fault then flattened into the bedding of lithotectonic unit 3. Movement along the thrust resulted in the formation of a superficial anticline similar to the Powell Valley anticline, but with a much narrower crest. When the rootless anticline ceased to move, the hanging wall broke so that shortening was distributed by the Rockwood and Chattanooga faults.

CHATTANOOGA - ROCKWOOD SYSTEM (SOUTHERN PART)

Recent drilling near Dayton, Tennessee has yielded some good shows of oil and a small amount of production from Mississippian limestones in the footwall of the Rockwood fault. One of the wells (Owensby No. 1A, section GG', Figure 7) penetrated the Rockwood thrust at 780 feet, entered fractured Pennington shales, and produced a little oil from the upper beds of underlying Monteagle Limestone (Mississippian). The Chattanooga Shale was reached at a depth of 1867 feet in the nearby (Peavyhouse No. 1) well. Based on the surface geology and projections of formation tops from the Gardner No. AA 9 well, Mississippian limestones beneath the Cumberland Escarpment and Rockwood thrust block are structurally high, and constitute the Cumberland Escarpment anticline (Swingle, 1961). In this area, the Cumberland Escarpment anticline is interpreted to be a small rootless fold that formed where a decollement in Silurian-Devonian shales cut upward across Mississippian limestones and flattened in the bedding of Mississippian and Pennsylvanian shales.

Several wells (Lavender No. 1, Shelton No. 1) were drilled in and near a tightly down-folded knot of Pennsylvanian strata in the Spring City window, partly because surface dips make it appear that the structure is anticlinal. The wells penetrated a normal stratigraphic sequence from the Pennsylvanian to the Chattanooga Shale and several had good shows of oil (Figure 7, section FF'). The main anticlinal trend, however, is somewhat to the west of the window, and probably is a reflection of a level change of the upper décollement from Silurian to Devonian to Mississippian and Pennsylvanian beds.

The Rhea Springs and the Post Oak structures, between cross sections EE' and FF' were thought to be windows by Rodgers (1950) and Swingle (1961) and his students. In general, they interpreted these steeply dipping, vertical to overturned beds of Ordovician to Mississippian age to represent the upturned limb of a synclinal fold that lay along and adjacent to the Chattanooga fault. These structures may well be windows, but I regard them as rootless fault-slices that are bottomed at shallow depths by the Rockwood fault and then probably by

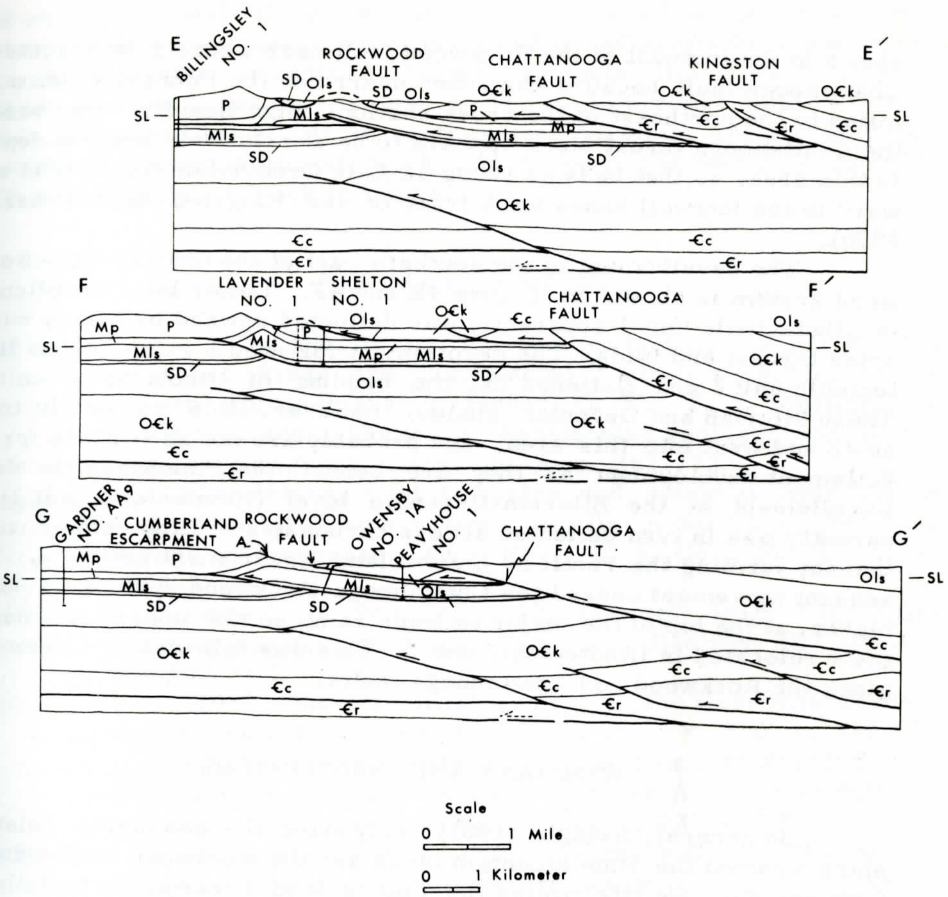

Figure 7. Cross sections of the southern part of the Chattanooga - Rockwood fault system, between Dayton and Harriman, Tennessee. Symbols explained in Figure 2.

subhorizontal beds of Mississippian age, similar to the structure of the Harriman Corner (cross section EE', Figure 7).

Cross section EE' is controlled at each end, on the northwest by the Billingsley No. 1 well and on the southeast by the seismic cross-section. Mississippian limestones in the Harriman window are structurally high and the anticlinal structure apparently results from a décollement level change from Silurian and Devonian to Mississippian and Pennsylvanian shales. Northwest translation along this upper level fault appears to be of greater magnitude in the Harriman Corner than along much of the Cumberland Escarpment anticline to the south, and is sufficient to produce almost two hundred feet of duplication within the Pennington Formation in the Billingsley well. Pennsylvanian beds are shown in the footwall of the Rockwood fault near where it intersects the Chattanooga fault because the latter overrides the Pennsylvanian a few miles to the northeast of the cross section. Significantly, the base of the Chattanooga thrust block appears to be at relatively shallow depths in this area, so that beds as young as Late Ordovician may extend eastward in the footwall beneath the trace of the Kingston fault (Harris, 1976).

The development of the southern part of the Chattanooga - Rockwood system is shown in Figure 4E and 4F. Lower level décollement in lithotectonic unit 1 apparently was deflected upward by deeply buried splay thrusts and folds. The décollement followed a ramp across lithotectonic unit 2 and flattened into the bedding of lithotectonic unit 3. These Silurian and Devonian shales, however, thin regionally to the south and west into this area, and probably are not as suitable for décollement development as they are beneath the Pine Mountain block. Décollement at the Silurian-Devonian level (lithotectonic unit 3) apparently was in turn deflected across carbonates of lithotectonic unit 4, thereby forming the rootless Cumberland Escarpment anticline. Subsequent movement caused the Chattanooga-Rockwood system to break higher, at the top of the major tectonic ramp so that upper level décollement relocated in lithotectonic unit 5. This was followed by imbrication along the Rockwood and Chattanooga faults.

SUMMARY AND CONCLUSIONS

In general, Swingle (1961) recognized the geometric relationships between the Pine Mountain block and the Rockwood - Chattanooga fault system. He interpreted the Cumberland Escarpment anticline as the result of a change in the level of décollement from Silurian and Devonian shales to Mississippian and Pennsylvanian shales (Swingle, 1960c). This interpretation seems to be correct between the Harriman Corner and Dayton, but north of the Harriman Corner anticlinal trends along the Allegheny front reflect a décollement level change from Cambrian to Silurian and Devonian shales.

Furthermore, Swingle (1961, p. 23) speculated on the general age relationships of thrust along the Allegheny front and was the first to propose that, "... faults become progressively younger toward the surface and to the east, each having been deflected upward by anticlinal structures formed as faults lower in the section step to higher glide zones." Subsequently, I used part of this important concept to explain structural patterns observable throughout the Valley and Ridge (Milici, 1975), and Harris' study of a seismic cross section in the Valley and Ridge provided the first firm evidence that the faults in the basal décollement migrated upward as thrusting proceeded in the Valley and Ridge.

The area between the Chattanooga block and the northwest limbs of anticlinal trends along the eastern Cumberland Escarpment contains strata of Mississippian age that produced or had shows of oil and natural gas in the region. In addition, Middle Ordovician limestone equivalents of the Trenton Limestone, which is known to produce small but commercial quantities of oil in southwestern Virginia, are probably beneath the Rockwood thrust block and beneath the leading edge of the Chattanooga thrust block in Tennessee.

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THE COWEETA GROUP AND COWEETA SYNCLINE: MAJOR FEATURES OF THE NORTH CAROLINA-GEORGIA BLUE RIDGE

By

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ABSTRACT

The name Coweeta Group is proposed for a group of metasedimentary and possible metaigneous rocks which occur in the east-central Blue Ridge of North Carolina and Georgia and overlie the rocks of the Tallulah Falls Formation. The group is composed of three formations. The oldest is the Persimmon Creek Gneiss. This is overlain by the Coleman River Formation, then the Ridgepole Mountain Formation.

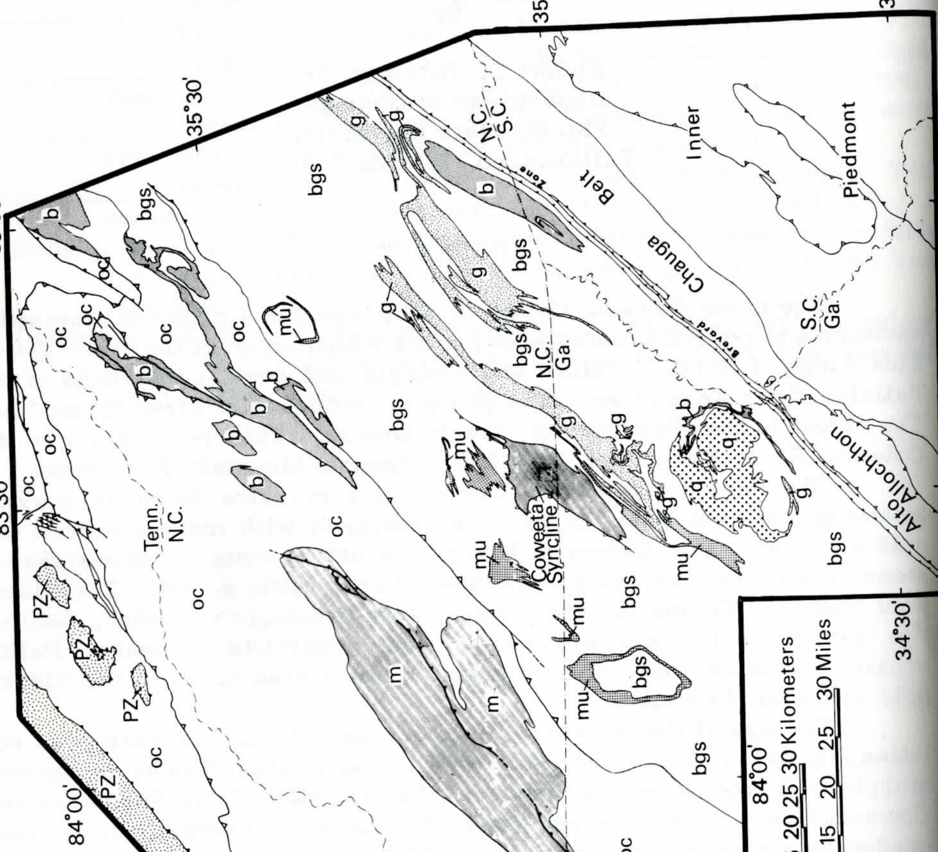
The Persimmon Creek Gneiss is a massive feldspar-quartz-biotite gneiss which is in part interlayered with metasandstone and pelitic schist. The Coleman River Formation consists of metasandstone, mafic poor gneiss (metaarkose?) and pelitic schist. The Ridgepole Mountain Formation has considerable lithologic variation, but biotite garnet schist and impure to clean quartzite dominate. Pelitic schist, metasandstone and metaconglomerate also occur in the Ridgepole Mountain Formation.

The age of the Coweeta Group is uncertain. An analogous relationship to the Ocoee and Chilhowee Groups, Great Smoky Group and Murphy belt rocks exists between the Tallulah Falls Formation and Coweeta Group, one of a sequence of unclean sedimentary rocks succeeded by a cleaner sequence. It could be a time equivalent unit to the Murphy belt sequence.

The Coweeta syncline is a structure resulting from a history of polyphase deformation. It appears to be overturned toward the east and the west limb is cut off by the pre-or synmetamorphic Shope Fork fault.

INTRODUCTION

The rocks of the Murphy belt have been known for many years as a younger metasedimentary sequence which overlies the late Precambrian Great Smoky Group in a structurally low position in the western Blue Ridge of Georgia and North Carolina. Another sequence of rocks, the Alligator Back Formation, occupying a position in the



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N.C.
S.C.
S.C.
Ga.
Coweeta Syncline
Brevard Zone
Chauga Belt
Inner Piedmont
Alto Allochthon
5 0 5 10 15 20 25 30 Kilometers
5 0 5 10 15 20 25 30 Miles

sequence similar to or near the base of the Murphy group was recognized by Rankin and others (1973) west of the Brevard zone along the North Carolina-Virginia border. The Alligator Back Formation likewise occupies a structurally low position.

The purpose of this paper is to describe an additional younger sequence of rocks which also occurs in a structural syncline. These rocks form a lithologically distinct sequence from the adjacent rocks and comprise a sequence of cleaner sediments therefore making it a coherent unit. This sequence, herein named the Coweeta Group, is located in the east-central Blue Ridge along the North Carolina-Georgia border (Figure 1).

Keith (1907, 1952) mapped the rocks of the Coweeta Group as Whiteside Granite, Roan Gneiss and Carolina Gneiss. Hadley and Nelson (1971) mapped these rocks as hornblende-biotite gneiss and biotite schist and gneiss of late Precambrian and Devonian age. This geologist first recognized this group of rocks during the reconnaissance mapping of Rabun and Habersham Counties, Georgia (Hatcher, 1971), but they were shown as Precambrian migmatitic biotite gneiss and granitic gneiss on the geologic map. Berry (1977; unpub. ms.) conducted a study of weathering rates on some of the Coweeta Group rocks. Several thin sections were made and metamorphic mineral assemblages were determined by Berry.

Acknowledgments

Support for early portions of the geologic field work which resulted in recognition of the Coweeta Group sequence was provided by the Georgia Geological Survey and National Science Foundation Grant GA-20321. Support for the detailed mapping of the Coweeta syncline in North Carolina has been provided by the North Carolina Geological Survey Section and the Tennessee Valley Authority. Cooperation of James E. Douglass, Wayne T. Swank and other staff members of the U. S. Forest Service Coweeta Hydrologic Laboratory greatly facilitated geologic mapping of most of the Coweeta syncline. Continued support for detailed geologic mapping has been provided by the North Carolina Geological Survey Section and by National Science Foundation Grant EAR76-15564.

Figure 1. Generalized geologic map of part of the southern Appalachians showing the location of the Coweeta syncline in relationship to nearby features. PZ-Paleozoic sedimentary rocks. oc-rocks of the Ocoee Supergroup. m-Murphy belt rocks. b-Precambrian basement rocks. mu-mafic and ultramafic rocks. bgs-biotite gneisses, schists, amphibolite, migmatite, minor granitic gneiss. q-quartzite. g-granitic gneisses.

The resolution of rock units of the Coweeta Group above the Persimmon Creek Gneiss was made by field assistant Louis L. Acker while mapping the Dillard Quadrangle, Georgia. His work, with slight modifications served as the basis for mapping of the major part of the Coweeta syncline in North Carolina.

A great deal is owed George D. Swingle who taught me the necessity of making systematic and detailed observations in carrying out geologic field investigations, yet to constantly place them into a regional framework.

Critical reviews by R. C. Milici, L. S. Wiener and C. E. Merschat are very much appreciated. However, I retain any responsibility for misinterpretations and other errors.

COWEETA GROUP

Introduction

The name Coweeta Group was first used informally by this geologist (Hatcher, 1974). The principal aim of this paper is to propose that the Coweeta Group and the formations making it up be formalized.

The Coweeta Group is named for the upper reaches of Coweeta Creek and its tributaries in the Coweeta Hydrologic Laboratory south of Franklin, North Carolina, the area where the largest outcrop of these rocks occurs in rocks of lower metamorphic grade. It is here that the large syncline preserving these rocks reaches its greatest width. However, just southwest of this area, all units in the sequence are exposed between the crest of Little Ridgepole Mountain and the west side of the valley of Bettys Creek to the east (southwest part of the Prentiss 7 1/2 minute quadrangle; Hatcher, in preparation).

Rocks of the Coweeta Group overlie the Tallulah Falls Formation. There is considerable contrast between the two units. While the Tallulah Falls Formation consists of metagraywacke, amphibolite, muscovite-biotite schist, lesser amounts of aluminous schist and minor quartzite, the Coweeta Group consists of larger amounts of feldspar-quartz-biotite gneiss, metasandstone and quartzite, aluminous schist, garnetiferous biotite schist and minor amphibolite. The contrast between the Tallulah Falls Formation and Coweeta Group is one of a sequence of unclean sediments and a sequence cleaner sediments.

Precise determination of the thickness of the Coweeta Group or of formations therein is impossible because of the nature and intensity of deformation affecting these rocks. The base of the Coweeta Group is known but its top is eroded and has therefore not been observed. The total thickness of the Coweeta Group is estimated to be roughly 2000 to 4000 m, but this estimate may be considerably in error.

The Coweeta Group has been subdivided into three formations: the basal Persimmon Creek Gneiss, overlain by the Coleman River

Formation, which is succeeded by the Ridgepole Mountain Formation. The sequence appears to be conformable and the units are related to each other by complex facies changes (Figure 2).

Persimmon Creek Gneiss

The Persimmon Creek Gneiss is named for exposures along Persimmon Creek in northwestern Rabun County, Georgia (Dillard, Georgia-North Carolina 7 1/2 minute quadrangle). Excellent exposures also exist along Patterson Creek in northwestern Rabun County and the flanks of Pickens Nose, Dryman Fork Creek, Rockhouse Knob and Bear-pen Creek and Mountain in southern Macon County, North Carolina.

The Persimmon Creek Gneiss directly overlies rocks of the Tallulah Falls Formation and underlies the Coleman River Formation. The Persimmon Creek Gneiss is the thickest unit in the Coweeta Group, making up one third to one half of its total thickness. The Persimmon Creek Gneiss was subdivided by L. L. Acker into upper and lower members in the Dillard Quadrangle, Georgia (L. L. Acker and R. D. Hatcher, Jr., unpub. data). The lower unit there consists of massive oligoclase-quartz-biotite (muscovite - clinozoisite / epidote - chlorite - garnet) gneiss containing little schist or metasandstone. The upper member consists of intricately interlayered gneiss of the same type as in the lower member along with quartz - oligoclase - biotite - (epidote-chlorite-garnet) metasandstone and biotite-muscovite-quartz (kyanite-oligoclase) schist. Such a twofold subdivision of this unit was found to be not resolvable farther north due to the greater complexity of deformation.

The Persimmon Creek Gneiss is easily identified by its coarse, even-grained texture (see Hatcher, 1974, Figure 6). It commonly forms massive exfoliated cliffs and ledges 3 m. or more high.

Coleman River Formation

The Coleman River Formation is named for exposures along the headwaters of Coleman River, Rabun County, Georgia. These rocks are also well exposed (and perhaps more accessible) along Ball Creek and Henson Creek in the Coweeta Hydrologic Laboratory in southern Macon County, North Carolina.

The Coleman River Formation consists of medium-grained metasandstone, metaarkose and quartz-feldspar gneiss with interlayers of coarse pelitic schist. The quartzofeldspathic layers are foliated and range up to 1 m. thick. The metasandstone and gneiss have a similar appearance but the metasandstone is higher in quartz and contains less feldspar. Both rock types consist of an assemblage of quartz-oligoclase-biotite-epidote-chlorite-(garnet-muscovite) metasandstone and gneiss, while the schist contains the assemblage biotite-muscovite-kyanite - staurolite - quartz - (garnet - oligoclase - epidote - iron oxides).

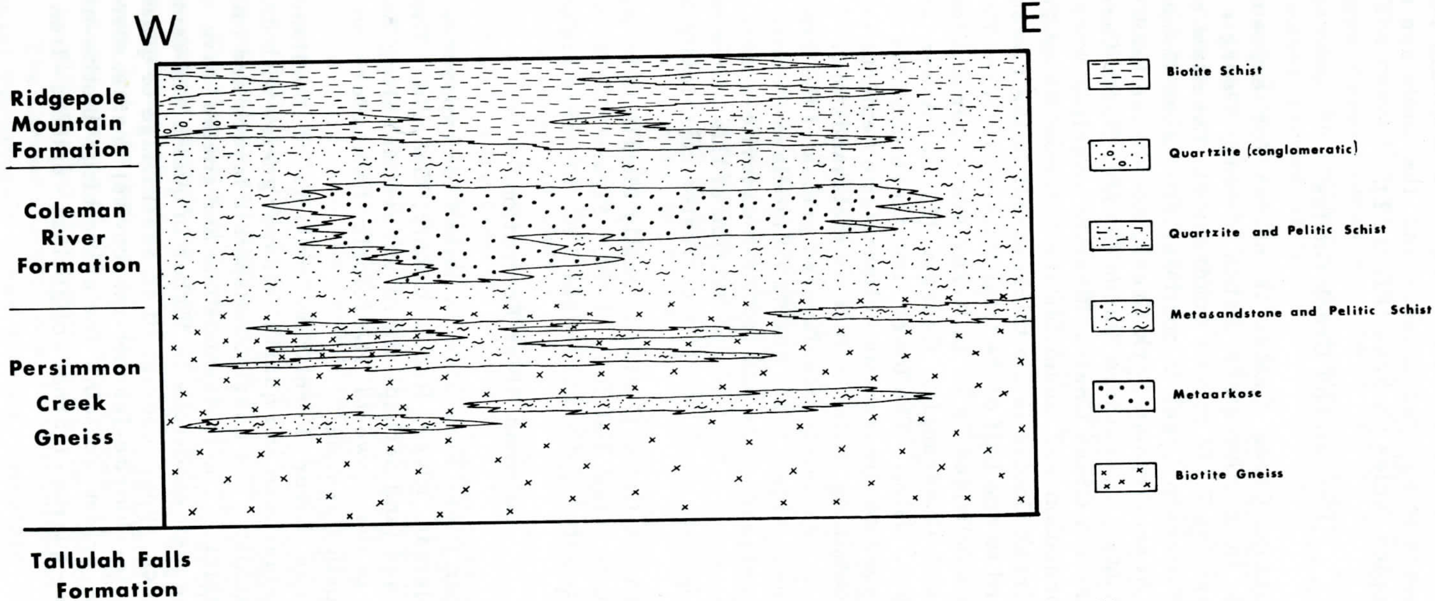

Figure 2. Possible facies relations in the Coweeta Group within the Coweeta Syncline.

Calcsilicate quartzite is also a minor but common constituent in this unit in some areas.

A structural-textural feature common in the Coleman River Formation is metasandstone possessing thin laminations or a "pin-striped" texture (see Hatcher, 1974, Figure 7). These laminations may be oriented parallel to the dominant foliation but commonly are not indicating that they represent an earlier S-surface which has not been transposed into what appears to be S_1. However, the laminations may be the product of initial transposition of bedding thereby making the present S_1 actually S_2 (very likely in light of my observations elsewhere in the Blue Ridge and Piedmont). Pinstriping of this type has been described in the Alligator Back Formation in the eastern Blue Ridge of northwestern North Carolina by Rankin and others (1973, p. 17). The pinstripped metasandstone also occurs in the overlying Ridgepole Mountain Formation, but less abundantly than in the Coleman River Formation.

Ridgepole Mountain Formation

The Ridgepole Mountain Formation, the uppermost unit of the Coweeta Group, is named from exposures on Ridgepole and Little Ridgepole Mountains, North Carolina, accessible by foot via the Appalachian Trail. More accessible exposures of this unit also exist along the road following Ball Creek in Coweeta Hydrologic Laboratory.

The Ridgepole Mountain Formation contains the greatest variety of lithologies ranging from pelitic schist to coarse biotite garnet schist to clean quartzite to muscovite-chlorite quartzite to garnetiferous metasandstone. The pelitic schist is similar to the aluminous schist in the Coleman River Formation. The biotite-garnet schist is also similar except for the dominance of biotite and flattened garnets which range up to 8 cm. in maximum diameter. The quartzites contain varying amounts of quartz, muscovite, chlorite, epidote, plagioclase and garnet. Some of the quartzites may have been conglomeratic. One specimen observed consists of quartz and muscovite in which each is separated into discrete flattened elliptical domains suggesting a protolith of mud matrix conglomerate.

Several lithologies of the Ridgepole Mountain Formation, e. g., biotite-garnet schist and clean quartzite, occur together in some areas. In other places, a single lithology dominates the outcrop belt of the unit. Some of the more impure sandstones and schists appear to be gradational into those of Coleman River Formation so that in places the boundary is difficult to ascertain.

The upper contact of the Ridgepole Mountain Formation has not been observed. It is assumed to be eroded.

Discussion

The rocks of the Coweeta Group comprise a distinctive sequence in the east-central Blue Ridge of North Carolina and Georgia. Within the group there are certain rock units, such as the Coleman River which are very uniform in character (metasandstone, metaarkose and schist) but at least one, the Ridgepole Mountain Formation is a very complex unit which grades into the unit below and contains several different facies (Figure 2).

The mineral composition and texture of the Persimmon Creek Gneiss suggest it is an orthogneiss. However, the interfingering and interlayering with metasedimentary rocks suggest a sedimentary origin. The Persimmon Creek Gneiss may once have been a fine-grained volcanic deposit which, due to its grain size, was more susceptible to a greater degree of recrystallization with resultant coarse texture and homogeneity. If it is of volcanic origin, the source of the volcanic material is unknown. It could also have been an intrusive which was emplaced near the metamorphic peak.

The age of the Coweeta Group is equivocal, like most of the other rock units of this portion of the Blue Ridge. The underlying Tallulah Falls Formation rests upon rocks that are likely 1100 m. y. old (Hatcher, 1976). The Tallulah Falls Formation is thought to be late Precambrian, possibly equivalent to part of the Ocoee Supergroup (Hatcher, 1971, 1973). The Coweeta Group could likewise be equivalent to part of the Ocoee and have a late Precambrian age. Its similar structural setting, relatively clean character and stratigraphic position suggest the possibility that it is equivalent to the Murphy belt rocks whose age is also equivocal (Hurst, 1955; Hadley, 1970; Power and Forrest, 1971; McLaughlin and Hathaway, 1973; Wiener, 1976). Whatever the age of the Coweeta Group, there is a repetitious trend among sedimentary rocks which immediately overlie basement to be impure and then succeeded by a cleaner sequence, reflecting changes in depositional environments. This may represent both lower relief source areas and filling (closing?) of a previously deep basin at the time of deposition of the Walden Creek and Chilhowee Groups, Murphy belt rocks and Coweeta Group. Another possibility is that these higher units formed by recycling of detritus (L. S. Wiener, written comm., 1977).

It appears likely that rocks belonging to the Coleman River and Ridgepole Mountain Formations occur farther west in the sillimanite zone in the central Blue Ridge. Detailed geologic mapping in Rainbow Springs and Shooting Creek quadrangles reveals a sequence and lithologies similar to these units (Hatcher, unpub. data). The Persimmon Creek Gneiss is either absent or has changed significantly in character.

Table 1. Mineral Assemblages in the Coweeta Group.

Unit/Lithology	Assemblage*
Ridgepole Mountain Formation	
Biotite-Garnet Schist	B M G C Q P
Quartzite	Q M C E G B St Mt
Pelitic Schist	M B G Ky C
Coleman River Formation	
Metasandstone	Q P M B E C G Mt
Quartz-Feldspar Gneiss	Q P M B G
Pelitic Schist	M B Q G Ky St Mt
Persimmon Creek Gneiss	
Biotite Gneiss	P Q B M E G C Ap Mt
Metasandstone	Q P M B E C G Mt
Pelitic Schist	M B Q G Ky St Mt

*Minerals listed in descending order of abundance.

Q - quartz
M - muscovite
B - biotite
G - garnet (probably almandine)
P - plagioclase (mostly oligoclase)
C - chlorite
Ky - kyanite
St - staurolite
E - epidote-clinozoisite
Mt - magnetite (or similar opaque)
Ap - apatite

METAMORPHISM

The rocks of the Coweeta Group have been metamorphosed to lower to middle amphibolite facies assemblages (Table 1). Pelitic (aluminous) schist contains both staurolite and kyanite. Both kyanite and staurolite are very fine-grained, have an irregular habit and appear to have been broken during deformation. Neither mineral is oriented parallel to S_1 foliation.

Garnets are euhedral to anhedral to skeletal (Figure 3). Many have a poikiloblastic texture while those in one sample have experienced more than one episode of growth. The latter have clear cores and poikiloblastic rims.

Chlorite occurs as a replacement mineral in many samples and appears to have developed as a retrogressive overprint. Epidote-clinozoisite replaces plagioclase in several sections probably resulting

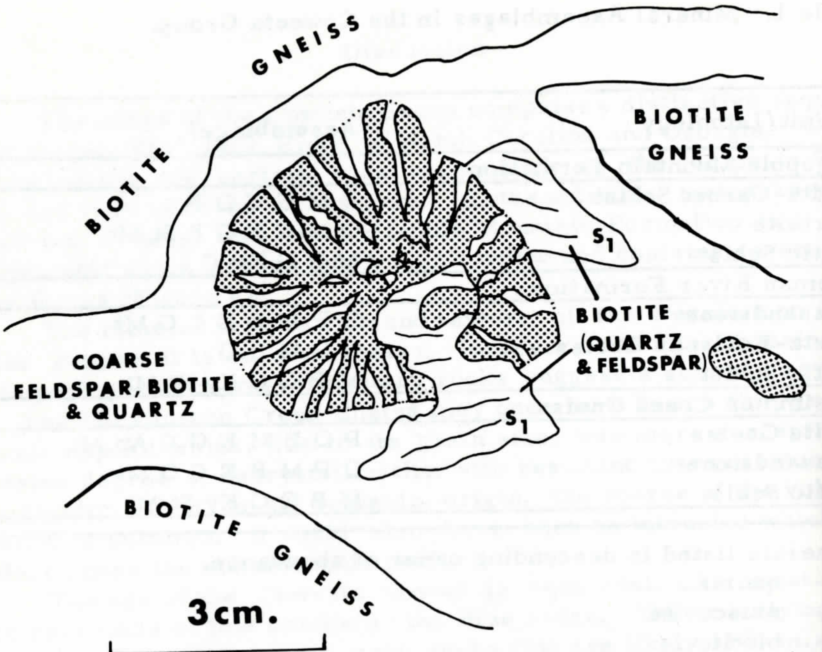

Figure 3. Garnet illustrating a common texture of garnets on all scales in Persimmon Creek Gneiss.

from the same retrogressive event. Late muscovite prophyroblasts which have grown across the S_1 foliation and do not appear at present to parallel any known S-surface may also have formed during this event.

COWEETA SYNCLINE

The Coweeta Group rocks are preserved in a structurally low area of the Blue Ridge of North Carolina and Georgia (Figure 1). They were first observed in a west-dipping series of rocks which appears to have been overturned toward the east. Large-scale earlier folds exist which have east-west and northeast axial trends. Their vergence is not known at present. Thus the Coweeta syncline is a polyphase structure (Figure 4) whose form was established during F_1 and F_2 folding (see Hatcher, 1976, Table 3; Hatcher, 1977, Table 1). F_1 folds are isoclinal and recumbent; F_2 folds are post-metamorphic upright folds formed under flowage conditions. Later folding has modified the structure only slightly on the macroscopic scale.

The northwest limb of the Coweeta syncline is cut off by the pre- or synmetamorphic Shope Fork fault (Hatcher, 1976). This fault was formed as a tectonic slide during F_1 folding and was refolded by F_2 and later folds (Figure 4). Its extent farther north and south is unknown.

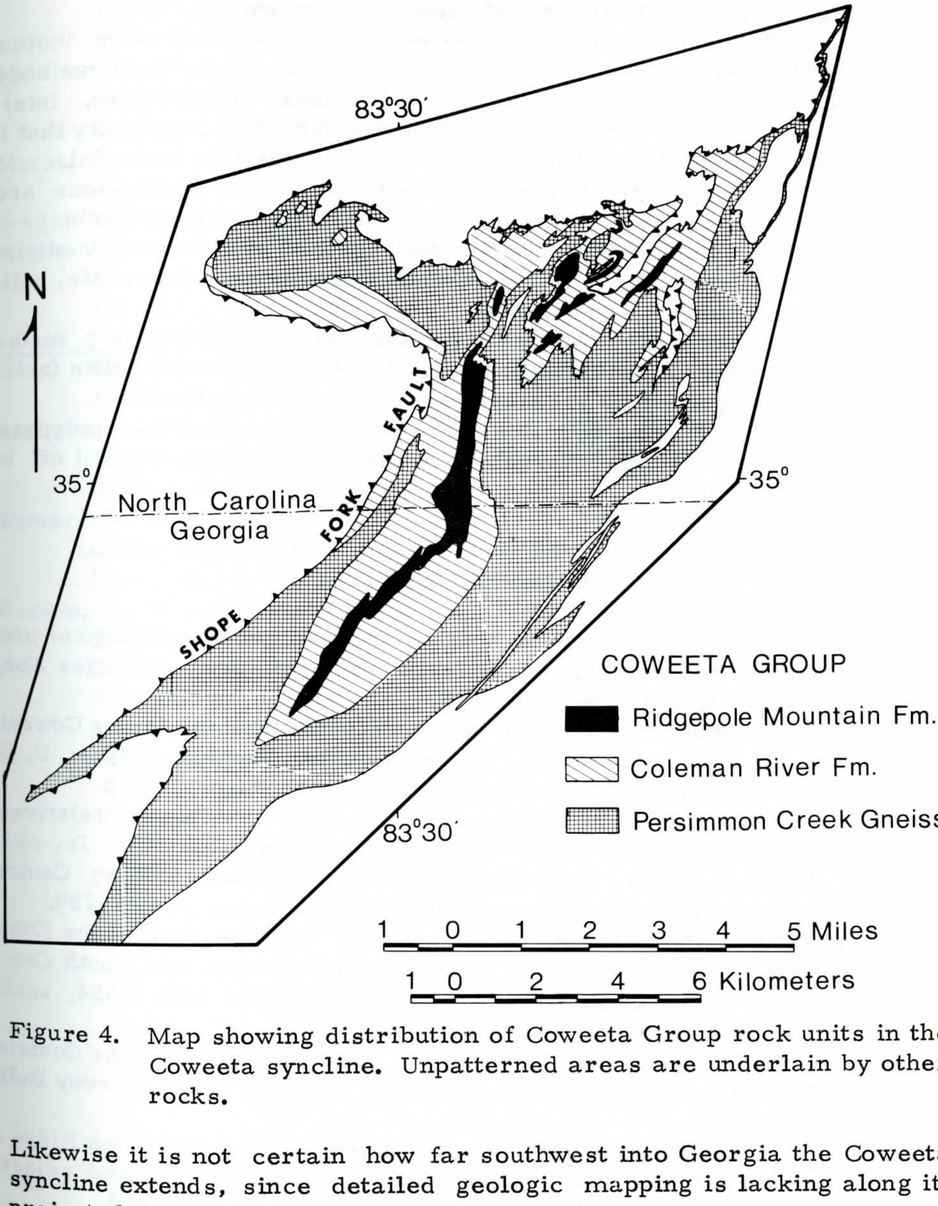

Figure 4. Map showing distribution of Coweeta Group rock units in the Coweeta syncline. Unpatterned areas are underlain by other rocks.

Likewise it is not certain how far southwest into Georgia the Coweeta syncline extends, since detailed geologic mapping is lacking along its projected trend.

CONCLUSIONS

1. The Coweeta Group consists of three formations overlying the Tallulah Falls Formation: the Persimmon Creek Gneiss, Coleman

River Formation and Ridgepole Mountain Formation.

2. The Persimmon Creek Gneiss is mostly massive biotite gneiss with interlayered metasandstone. It may be an intrusive body or a metasedimentary rock type but its textures, homogeneity, interlayering and lack of a contact aureole bring forth the possibility that it is derived from metamorphism of fine-grained volcanic material.

3. Coleman River and Ridgepole Mountain Formations are metasedimentary units. The former is a fairly uniform assemblage of pelitic schist, metasandstone and metaarkose. The latter is a complex assemblage of metaquartzite, metasandstone, metaconglomerate, pelitic schist and biotite garnet schist.

4. Rocks of the Coweeta syncline were progressively metamorphosed to the staurolite-kyanite subfacies of the amphibolite facies and later retrograded by greenschist facies metamorphism.

5. The Coweeta syncline is a structure resulting from polyphase deformation. It is east vergent and the northwest limb is cut off by the Shope Fork fault.

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STRUCTURE OF THE SLEEPING GIANTS RANGE, TALLADEGA COUNTY, ALABAMA

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ABSTRACT

The easternmost portion of the southern Appalachian Valley and Ridge Province in Alabama, known as the Coosa Valley, contains sedimentary rock units ranging in age from Cambrian to Pennsylvanian. Although the strata have been at most weakly metamorphosed, folding and faulting have produced a complex structural pattern in the area interpreted as resulting from either a series of thin, northwest transported imbricate thrust slices or a complexly folded nappe or large recumbent fold. In addition to the controversy about the structure, the age of certain stratigraphic units has been disputed. Detailed mapping of a critical portion of the Coosa Valley has provided information which may resolve these controversies.

A major problem of the Coosa Valley stratigraphy has been the lack of paleontological evidence for the age of certain clastic units. In particular, strata that were correlated with the Lower Cambrian Weisner, Shady, and Rome Formations by early workers on basis of lithologic similarity have been reevaluated by recent mapping, and a late (Devonian-Mississippian) Paleozoic age has been suggested by some workers. Recent discoveries of a middle Cambrian trilobite fauna and of archeocyathids in this area indicate the early workers were basically correct in their identification of the Weisner, Shady and Rome Formations in the Coosa Valley.

Bedding attitudes within the Weisner, and structural relationships observed between the Weisner-Shady sequences and surrounding younger strata indicate the Weisner-Shady sequences are thrust over younger strata by a nearly horizontal fault. The Weisner and Shady underlie north-to northeast-trending ridges which are apparently klippen, rooted either within or beneath the Talladega metamorphic belt that lies to the southeast.

INTRODUCTION

The southeastern part of the Appalachian Valley and Ridge Province in Alabama is a broad northeast-trending lowland called the Coosa

Valley. Within the Coosa Valley, formations of Paleozoic age are folded and faulted and locally show the effects of low-grade metamorphism. The Coosa Valley is bordered on the southeast side by the Talladega metamorphic belt, the northwesternmost belt of the Appalachian Piedmont Province.

The Cambrian-Ordovician Knox Group, consisting of dolostone and limestone, is the most widely distributed stratigraphic unit in the Coosa Valley. Other units, consisting of carbonates and clastics, both older and younger than the Knox, are also present. In the northwestern part of the Coosa Valley, fossils are sufficiently abundant to permit reliable stratigraphic interpretation. However, in the southeastern part, the increasing degree of penetrative deformation and metamorphism has apparently destroyed whatever fossils were present. Also, at least one formation appears to be fluviatile in origin and may never have possessed fossiliferous zones. Relative lack of fossils in the southeastern part of the valley has left the stratigraphy, and thus the structure, open to question. The Sleeping Giants, a northeast-trending ridge complex in the Coosa Valley, lies northwest of the city of Talladega. It is composed of shale, quartzite and chert, and is particularly critical to an understanding of the stratigraphy and structure in this region (Figure 1).

PREVIOUS WORK

Butts (1926, 55-62) described the geology of the Coosa Valley and interpreted the stratigraphy on the basis of fossils, lithologic similarity, and apparent stratigraphic sequence. Interpretation of the Weisner Formation (Lower Cambrian) as the oldest unit in the valley was based on fossils obtained from strata called Weisner in Georgia. Butts described the Lower Cambrian Shady Formation as a carbonate unit possibly 517 feet thick (well data) overlying the Weisner and overlain by the Lower Cambrian Rome Formation. He stated that the Lower Cambrian age could be inferred also from the presence of <u>Salterella</u> and <u>Archeocyathus</u>, found in the Shady in Georgia. Butts identified the Lower Cambrian Rome Formation and Middle Cambrian Conasauga Formation in the Coosa Valley on the basis of fossils and lithologic characteristics. He described the Rome as green and maroon shale and thick beds of hard, fine-grained, ferruginous and calcareous sandstone. He mentioned fossiliferous yellow clay beds which are remnants of leached limestone beds. Butts described the Conasauga of the Coosa Valley as one third limestone and dolomite and two thirds shale. He reported a 100-foot thick interval of thick bedded, dark gray limestone and some flaggy limestone in and near the city of Talladega as being Conasauga, and described an inarticulate brachiopod, <u>Obolus</u>, from a locality at the base of the Conasauga three miles west of Talladega.

Butts described three carbonate formations; in ascending order,

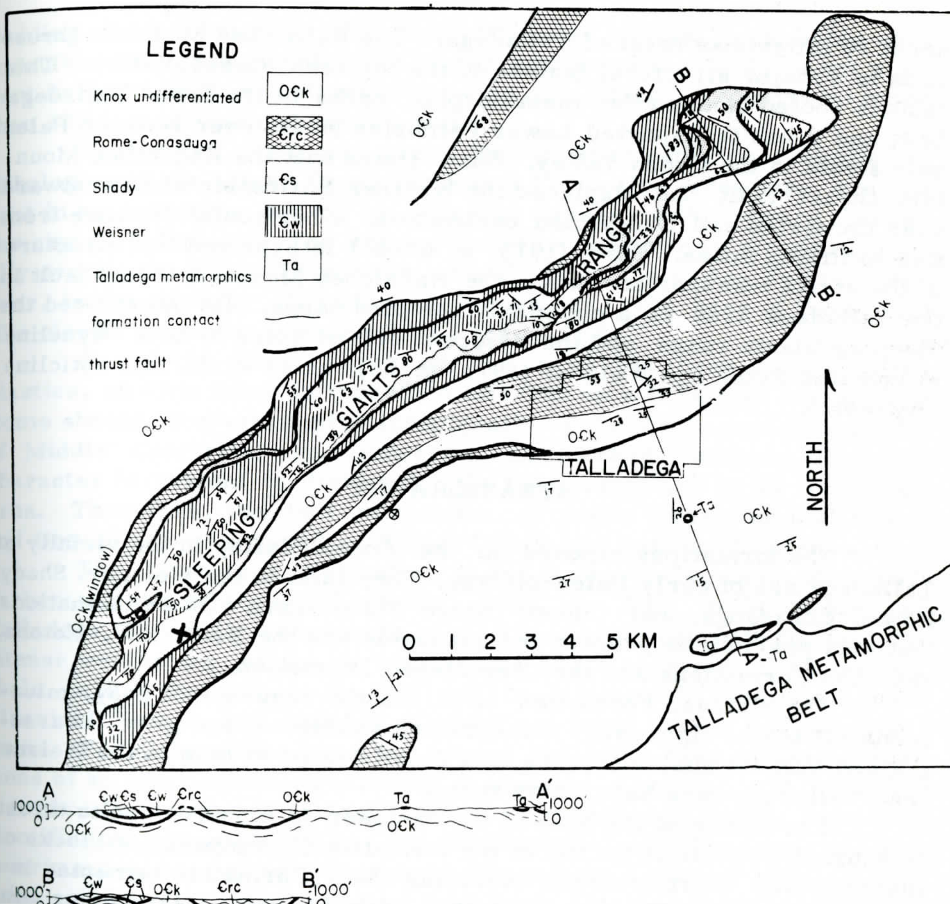

Figure 1. Geologic map and structural cross-sections of Sleeping Giants Range and environs, Talladega County, Alabama. Archaeocyathid site at west end of Sleeping Giants shown by X.

the Brierfield, Ketona, and Bibb Formations, all dolomites of Late Cambrian age, exposed near Montevallo in Shelby County, Alabama. Butts (1926, p. 84) suggested the possibility that these formations are present in the southeasternmost Coosa Valley near Talladega. Butts reported the Upper Cambrian Copper Ridge Dolomite as the most widespread formation in the Coosa Valley. He stated that dense, hard, jagged chert, a weathering replacement product on the Copper Ridge Dolomite, is the most distinctive feature of the formation.

Shaw (1970) described the goology in the Coosa Valley and adjacent Talladega metamorphic belt in the Sylacauga-Childersburg region, about 20 miles southwest of Talladega. The Kahatchee Mountain thrust fault is a major structural feature of the eastern Coosa Valley. This fault is rooted beneath the metamorphic rocks of the lower Talladega belt, and it has transported Lower Cambrian strata over younger Paleozoic strata in the Coosa Valley. Shaw stated that the Kahatchee Mountain thrust fault had displaced the Weisner Formation northwestward over Cambrian and Ordovician carbonates, a horizontal distance from five to fifteen miles. Shaw (1973, p. 60-62) interpreted the structure of the strata transported above the Kahatchee Mountain thrust fault in the Talladega area as a complexly refolded nappe. He interpreted the Sleeping Giants range and the adjacent belt of Rome to be a "syncline in inverted Rome and Weisner (1973, p. 60), or an synformal anticline (my term).

STRATIGRAPHY

The formations exposed in the Coosa Valley in the vicinity of Talladega are of early Paleozoic age. They include the Weisner, Shady, Rome, Conasauga, and Copper Ridge Formations. Other formations that may possibly be present and mappable are the Brierfield, Ketona, and Bibb Formations and the Chepultepec Formation.

The Weisner Formation is a clastic sequence of gray, micaceous siltstone, light grey, fine-grained sandstone; and grey, coarse-grained arkosic quartzite. The only fossils found to date in the Weisner near Talladega have been worm tubes.

The nature of the base of the Weisner is unknown due to thrust faulting. A breccia at the top of the formation is composed of blocks of quartzite and chert from the overlying Shady Formation cemented in a sandy, limonitic, siliceous matrix. Lack of shear features or drag folds in shaly partings near the upper contact suggests the breccia is probably not of tectonic origin. I have interpreted the breccia as the result of karst development post-dating deposition of the Shady Formation.

The Shady Formation in the Talladega area is represented by massive rounded ledges and boulder of brown, coarse, cellular chert locally containing abundant coarse quartz sand grains and yielding archaeocyathids at one locality (Figure 1). The Shady also contains intervals of coarse-grained, limonitic quartzite up to several feet thick. The occurrence of <u>Archeocyathus</u> (Bearce & McKinney, 1978) is evidence of Early Cambrian age and serves as verification that this chert does in fact represent the Shady.

The Weisner and Shady formations are restricted in occurrence to the Sleeping Giants range and its apparent extensions. The close and restricted association of these two formations indicates a stratigraphic relationship. Generally, Shady chert in the Sleeping Giants overlies Weisner strata in normal succession as indicated by cross-bedding in the Weisner. The Shady/Weisner contact is probably conformable and gradational. The upper contact of the Shady is not present in the vicinity of Talladega because of faulting. Probably no more than 100 feet of Shady chert is present.

Variegated shale and grey, fine-grained sandstone, similar to lithologies of the Lower Cambrian Rome Formation elsewhere, are present in an extensive belt of clastics that passes northeastward through the city of Talladega. The Rome Formation is known to stratigraphically overlie the Shady Formation elsewhere in the Coosa Valley; however, in the vicinity of Talladega it does not. The base of the Rome-like clastic sequence is marked by a thrust fault.

Fossils of Early Cambrian age have not been found in these clastics, and the Rome is therefore inferred on a lithologic basis. The Rome should be overlain stratigraphically by the Conasauga Formation of Middle Cambrian age. However, no marked change in lithologic character is evident to distinguish the two formations in the Talladega area. Therefore, the belt of clastics, probably containing both Rome and Conasauga, is mapped as Rome-Conasauga undifferentiated.

The Conasauga Formation in the Coosa Valley has a clastics/carbonate ratio of about 2/1 (Butts, 1926, p. 70). In the vicinity of Talladega, the high percentage of clastics, combined with the fact the intense weathering of the carbonates has left a residuum of variegated shale, makes the Conasauga difficult to distinguish from the Rome. Discoidal lenses and nodules of chert are found locally with the residuum. Fossil localities near the city of Talladega have yielded specimens of an inarticulate brachiopod, possible Obolus, and trilobite carapace fragments. Abundant trilobite cephalons and pygidii have been identified as Glossopleura sp., an "Ehmaniella"-like ptychoparioid and a corynexochid trilobite. This fauna represents a medial Middle Cambrian age; similar fossil species have been obtained from beds referred to as Rutledge (second from the bottom of a six-fold formational sequence forming the central belt of the Conasauga Group in Tennessee) elsewhere in Alabama and in Virginia (Palmer, 1974 written commun.). The Conasauga clastics and residuum are overlain, apparently conformably, by dolomite beds. Thickness of the Rome-Conasauga sequence cannot be measured, since the entire sequence is not present.

The carbonate sequence overlying the Rome-Conasauga has been mapped as Knox undifferentiated. This carbonate sequence occupies a larger portion of the Coosa Valley than any other unit. Attempts at subdividing it have met with difficulty because of the general lack of fresh carbonate exposures and ground water replacement of the original carbonate by a chert which has a fairly constant physical character over the entire Coosa Valley.

The lowest beds of the carbonate sequence overlying the Rome-Conasauga clastics are medium to light gray, with a dark grey or maroon mottling, coarsely crystalline, massively bedded dolostone with discontinuous undulatory laminations, small scale cross laminations, and maroon and black shale partings. Cryptozoan structures up to 1 foot in width and height are also common. Chert, replacing these lowermost beds, resembles that representing the remainder of the Knox. These lowermost carbonate beds may correlate with the Brierfield, Ketona, and Bibb Formations of Shelby County to the west of Talladega (Butts, 1926, p. 81-84).

Dolostone higher in the Knox carbonate section is generally light grey and finely crystalline with lenses of dark grey chert. Light grey to amber, dense, tough, jagged, massive chert and soft, light grey, mealy cavernous chert, ground water replacements of the original dolomite, are both abundantly present. Intervals of fine grained, tan, cross-bedded sandstone up to 30 feet thick also occur in the Knox, and are useful in determining bedding tops.

STRUCTURE

Three structural models have proposed for the Coosa Valley in the vicinity of Talladega. Butts (1926) depicted the structure as being controlled by imbricate thrust faulting. Butts mapped the strata of the Sleeping Giants and other similar ranges in the eastern Coosa Valley as being bordered to the northwest by a thrust fault and as being stratigraphically contiguous with younger formations to the southeast.

Shaw (1973, p. 60-62) suggested that the Sleeping Giants strata and the belts of Rome-Conasauga that partially border and lie adjacent to the Sleeping Giants are all part of a major complexly folded recumbent fold, or nappe, transported northwestward on a thrust fault that is rooted beneath the metamorphosed sediments of the Talladega Belt (Figure 2). Shaw depicted the Sleeping Giants as the inverted nose or crest of an anticline within the nappe. Implicit in Shaw's model is a high degree of plastic deformation.

My field studies indicate that the Sleeping Giants range is a klippe, in faulted contact with the belts of Rome-Conasauga and Knox in the surrounding lowland. Detailed mapping of the Sleeping Giants and surrounding areas has revealed the following features shown in Figure 1:

1. Strata within the Sleeping Giants are practically universally facing upward; bedding tops face predominantly to the southeast. This is in direct conflict with Shaw's assertion that the strata on the northwest side of the Sleeping Giants are inverted, an essential condition for his inverted anticline.

2. Folds within the Sleeping Giants of amplitude larger than outcrop or road cut dimensions are of a broad open nature. Pronounced closure is found only at the northeast end of the Sleeping Giants (Figure 1). The most notable structural characteristic of the Sleeping Giants is the imbrication shown especially in the northeastern part of the range. This suggests that deformation was of a brittle, rather than a

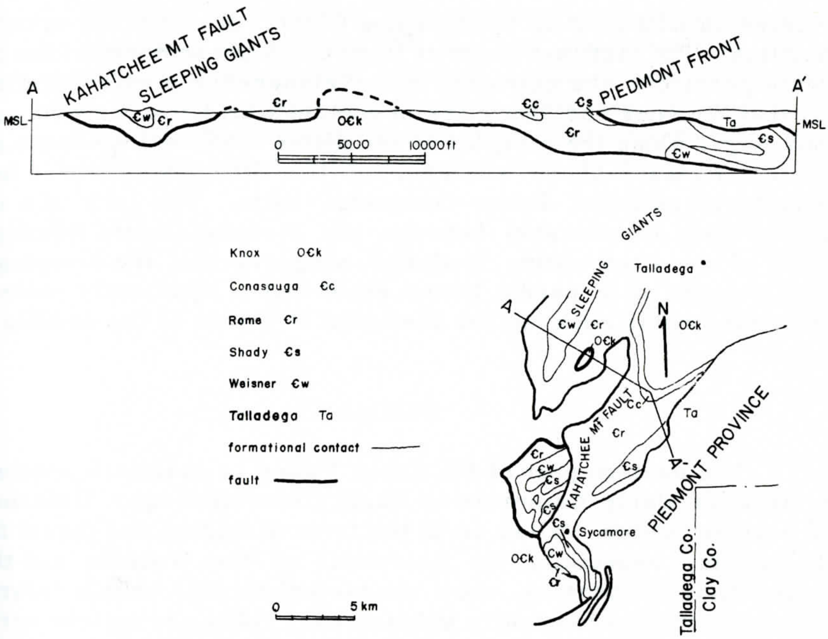

Figure 2. Geologic map of Sleeping Giants and vicinity, after Shaw (1976, p. 408). Structural cross-section after Shaw (1973, p. 62). Section line A-A' approximated by Bearce, due to lack of section line in Shaw (1973, Figure 11).

plastic nature. The individual thrust slices in the northeastern part of the range have subsequently been folded together into an eastward-plunging anticline. (Sec. B-B', Figure 1).

3. The Sleeping Giants are bordered and directly underlain by Knox carbonates as well as by Rome-Conasauga. Knox chert is exposed in a small window near the southwest end of the Sleeping Giants (Figure 1). Cryptozoan-bearing Knox dolomite occurs on the southeast side of the Sleeping Giants at Talladega.

4. A large part of the lowland region immediately surrounding the Sleeping Giants range is underlain by the Knox carbonate sequence. Detailed mapping in selected parts of this region indicates that the Knox has been folded into northeast-trending upright open folds. The Rome-Conasauga has been thrust faulted onto Knox. Bedding in the Knox has shallow to moderate dip; cryptozoan structures and cross-bedding indicate that Knox beds face upward throughout the lowland region surrounding the Sleeping Giants.

Shaw postulates that the Weisner and Rome (Rome-Conasauga in this paper) are stratigraphically contiguous in the Sleeping Giants. The intervening Shady is absent due to erosion or non-deposition. Shaw's inverted anticline within the Sleeping Giants relies heavily upon this assumption. The fact that Shady is very much present within the Sleeping Giants precludes the existence of a Weisner-Rome unconformity. The fact that Weisner and Shady are in contact with Knox, as well as Rome-Conasauga, along the margins of and within the Sleeping Giants suggests that the Weisner-Shady sequence of the Sleeping Giants is in faulted contact with adjacent Rome-Conasauga belts. The lack of a uniform stratigraphic relationship between the Sleeping Giants lithologies and strata of the surrounding lowlands suggests that the Sleeping Giants form a klippe of a parent thrust sheet that is apparently rooted within the metamorphic rocks of the Piedmont Province to the southeast.

SUMMARY

The eastern part of the Coosa Valley is underlain predominantly by strata of Early Cambrian to Early Ordovician age. Deformation in the eastern Coosa Valley is in the form of folding and thrust faulting. The strata, especially the quartzites of the Weisner and the thick bedded Knox carbonates, show characteristics of brittle deformation. Thrust sheets are internally imbricated. Folds are mainly upright and open. An upright attitude prevails throughout the region examined, as shown by cross-bedding and cryptozoan structure orientations.

Shaw (1973) has postulated that the Weisner and Rome are present in the eastern Coosa Valley as part of a nappe, resting on Knox. Field evidence in the Talladega area does not support this concept. Instead, the Weisner and Shady are present as klippen of a thrust sheet rooted within the Talladega metamorphic belt to the southeast. These klippen rest on a footwall of folded and thrust faulted Rome-Conasauga equivalents and Knox.

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THE IMPACT OF ACTIVE COAL MINING AND ORPHAN COAL MINES ON THE WATER QUALITY IN THE SOUTHERN COALFIELD OF EAST TENNESSEE - A COMPARISON

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ABSTRACT

This report is an attempt to emphasize the severity of environmental impact of coal mine drainage from mines abandoned prior to strip mine reclamation laws. Water quality in two areas is compared; an area of active strip mining and an area impacted by orphan mines. The depression of water quality below background values of the measured parameters (pH, specific conductance, and sulfate concentration) is nearly the same for an area with drainage from mines in a condition no different than when they were abandoned 20 to 30 years ago as for an area receiving drainage from an active strip mine.

INTRODUCTION

The addage that time is a cure for all ills has often been extended to include mitigation of lands impoverished by past mining activities. The assumption of this point of view is, then, that natural processes given ample time should be able to restore a riparion system impacted by mining to near pre-mining conditions. However, the enigma of this idea is just what is an ample length of time, and what conditions can be expected while waiting for time to heal the wounds? State mine reclamation laws and, more recently, federal reclamation laws have greatly reduced the potential for future long-term impacts associated with strip mining; however, regardless how stringent mining laws may become, total elimination of environmental stresses attendent to mining operations is virtually an impossibility. Thus, one might interpret the current regulations as being adequate to reduce the intensity of immediate environmental stresses, and perhaps eliminating the long-term stresses associated with strip mining for coal.

This paper attempts to point out that natural recovery of orphan mines, at least on the Cumberland Plateau in Tennessee, has been so

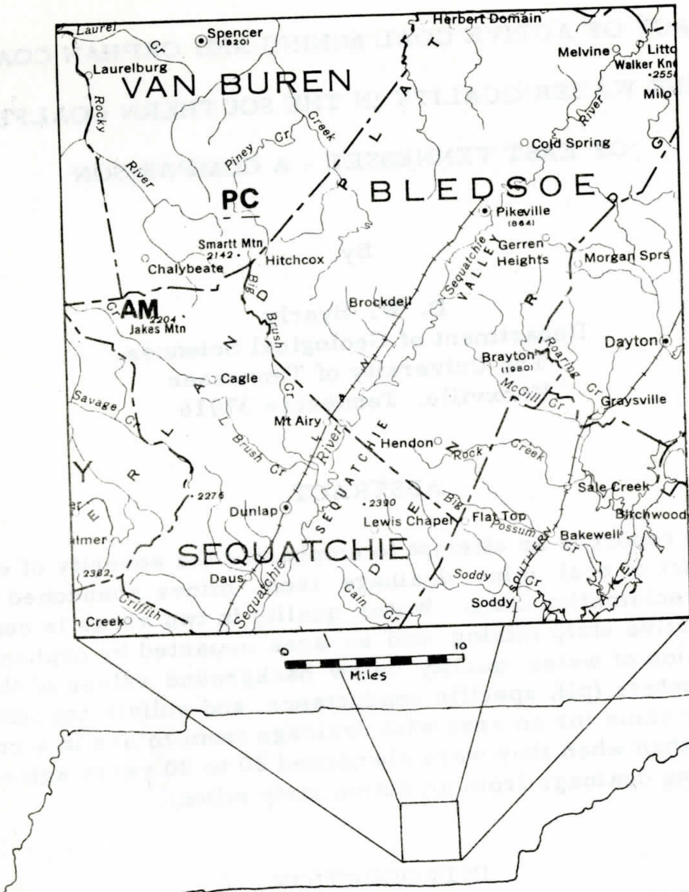

Figure 1. Index map showing the study areas. AM is the area of active mining and PC is the area of orphan mines.

slow that the present impact of abandoned mines that are 20 to 30 years old is as deleterious to the environment as the immediate effects of active mining. In order to demonstrate this, water quality data from an active strip mine operation is compared with water quality from a watershed pocked with orphan mines with no active mining. Both areas are situated in the southern coalfield of East Tennessee approximately 10 miles apart in Sequatchie and Van Buren Counties (Figure 1).

The stratigraphy, climate, soils, and vegetation are virtually the same in both areas, with the variables being time and mining methods. In the case of the area where orphan mine impact was evaluated, only natural recovery had been at work on a land degraded by irresponsible mine operators who were not restricted by mining laws in any way. The active mine operation was evaluated to determine how effectively a mine operator was able to abate pollution under the constraints of law (Office of Surface Mining, 1977; U. S. Public Health Service, 1962).

Background water quality values were determined for both areas by analyzing water samples collected from stream points above the mined areas. Two additional sampling points within each of the areas provided the data used to evaluate the impacts of the mining.

Acknowledgments

Portions of the data presented in this report are the results of studies sponsored by Argonne National Laboratory (Contract No. 31-109-38-3445) and the Tennessee Department of Conservation. Also, credit must be given to the Tennessee Valley Authority and the Tennessee Division of Water Quality Control for their support in the investigations. Gale Claytor and Clark Fletcher, former graduate students at the University of Tennessee, aided in gathering and analyzing the data in the above studies.

GEOLOGIC SETTING

Both study sites are situated on the rolling upland of the southern Cumberland Plateau in Tennessee. The lithogenesis of Carboniferous strata capping the Plateau has been attributed to a southward prograding delta (Milici, 1974). The Sewanee Conglomerate, Whitwell Shale, and Newton Sandstone, lower Pennsylvanian strata which comprise the areal geology in the two study areas, are interpreted as representing barrier foreshores and barrier-back lagoons of this prograding delta (Ferm and others, 1972). The Sewanee, mainly a conglomeratic orthoquartzite, ranges from 50 to 100 feet in thickness, and probably represents the barrier foreshore behind which the muds and organic matter comprising the superjacent Whitwell sediments accumulated. The Whitwell Shale is the main coal-bearing unit in the region and represents the barrier-back lagoonal environment. The Whitwell ranges in thickness from 69 to 90 feet, and although mainly a carbonaceous shale, it is locally silty and high in iron disulfides. A prominent fine-grained subgraywacke usually high in iron dissulfides, locally separates the Richland Coal which occurs in the basal 20 feet of the Whitwell from the Sewanee Coal. The Sewanee seam is the primary coal mined in the southern coalfield. The Newton Sandstone above the Whitwell ranges from 80 to 100 feet in thickness and can be classed as a fine- to medium-grained subgraywacke. The Newton possibly represents the filling of the lagoonal marsh during progradation.

Currucio and Ferm (1974) have demonstrated that rocks representing transitional paleoenvironments of their model such as those described above bear sulfide forms that are highly susceptible to generating

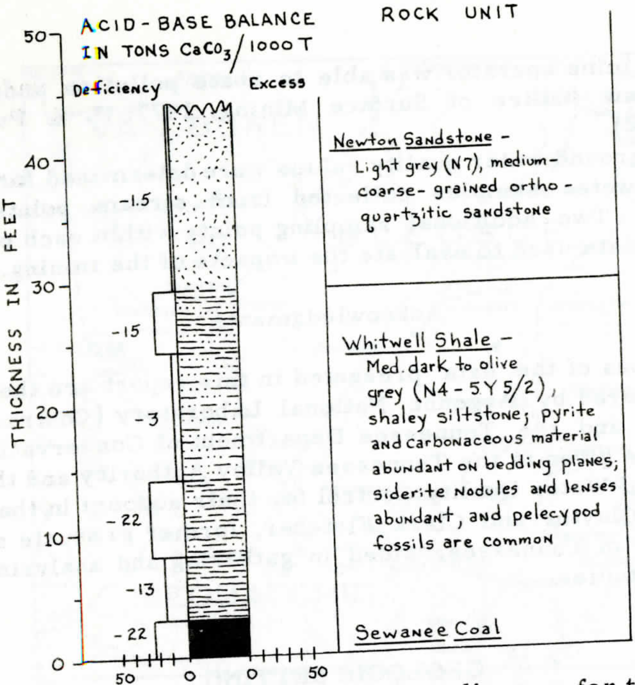

Figure 2. An acid-base balance diagram for the strata exposed in the highwall of the active mine.

acid conditions upon weathering. Figure 2 is an acid potential diagram based upon analyses of the Sewanee Coal, Whitwell Shale, and Newton Sandstone of the highwall in the active mine. The absence of carbonate material other than scattered siderite nodules is evident in the low neutralization potential values for these rocks.

WATER DATA

Water data were compiled from monthly sampling of streams in both areas. The sampling in the orphan mine drainage was done in 1974-75 in order to establish baseline data for a reclamation project, and the sampling at the active mine was done during 1976-77. Meterological conditions during the sampling periods are reflected in the TVA rainfall data shown in Figure 3. These data show the similarity in conditions during the two years, but also the diversity in the meteorological conditions throughout each sampling period. Collecting was done in dry weather as well as during the after precipitation events. The streams in both areas are low volume streams, but the stream flows in the Piney Creek drainage were on the average about 1.5 order of magnitude

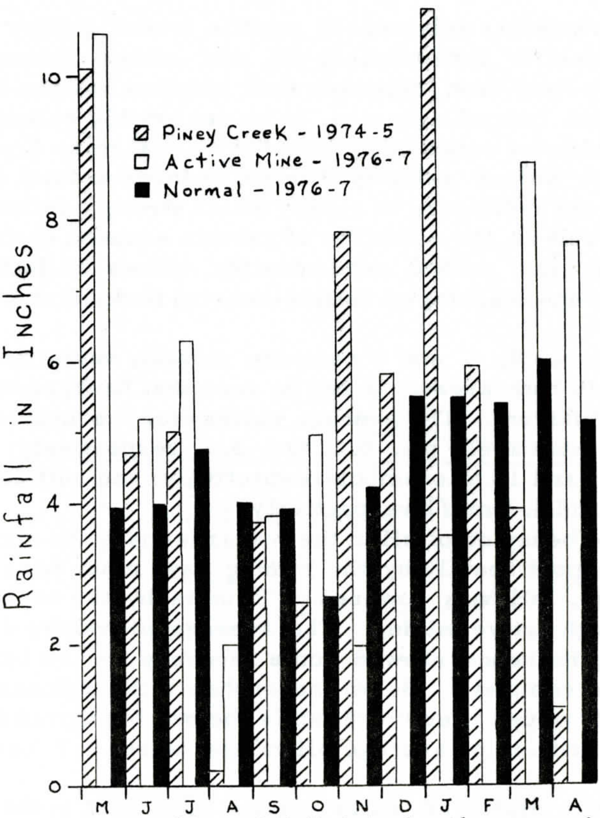

Figure 3. Monthly rainfall data for the years in which the studies were conducted. (Tennessee Valley Authority, 1974-77).

greater than those at the active mine. The seasonal fluctuations, however, were relatively the same, following the trends of the precipitation events.

Samples designated as PC-1 and AM-1 represent background water quality values for Piney Creek (the orphan mine area) and the active mine, respectively. Other sample points in the two areas include: PC-2, about the mid-point on Piney Creek, in the midst of numerous small tributaries discharging orphan mine effluent into Piney Creek. This point has been selected to represent the bulk quality of orphan mine discharge. Sample locality PC-3 is located downstream from PC-2 near the mouth of Piney Creek. AM-2 represents active mine discharge, and AM-3 is the water quality sample point below the confluence of AM-1 and AM-2. Despite the differences in the magnitudes of the flow distances and the water volumes between the two areas being compared, I believe some rational conclusions may be drawn when the water quality data are compared on a relative basis.

Parameters selected to profile water quality in the two areas include: specific conductance, pH, and sulfate concentration. These parameters have been demonstrated by other studies including, several by state and federal agencies, to be reasonably reliable indices of the overall quality of water being effected by coal mine drainage. Specific conductance values reliably indicate changes in total dissolved solids; pH values are indicators of acidity which strongly influences the aquatic biota, as well as the solubility of certain common metals like iron and manganese; and sulfate concentration values reflect the increase in weathering and leaching of sulfide-bearing rocks disturbed in the course of mining.

Figures 4, 5, and 6 compare monthly variations in parameters measured in both areas. It can be seen that background values are not radically different. The average values for the orphan mine area and the active mine are: pH: 6.1 and 5.1 respectively; conductivity: 40 micromhos and 16 micromhos respectively; and sulfate concentrations: 7.1 mg/l and 4.1 mg/l respectively.

The behavior of pH in the two areas is quite similar. The points influenced most heavily by the mining have much lower pH values. On the average, pH was lowered 1.8 units by the orphan mines and 1.6 units through active mining. The average pH values downstream from the mining influence showed some recovery toward background levels, but was not complete. Below the orphan mine influence Piney Creek's average pH values were 0.8 units below background and downstream from the active mine average pH values were 0.7 units below background.

Total dissolved solid values measured in the studies in mg/l were consistently around 67% of the conductivity values. Mine influences as recorded at PC-2 and AM-2 were dramatically higher than background values. PC-2 averaged 793% higher than background, and AM-2 was 3794% higher. The rather large difference between the increases could be attributed to the difference in stream sizes, but it is also probably due to the fact that AM-2 represents active mining. Through dilution the downstream values for conductivity are reduced to the point that they are 270% and 513% respectively above background levels for Piney Creek and the active mine respectively. A residual effect is observable downstream in both cases, and that of the orphan mine area is not much less than that of the active mine.

A rather great difference can be observed in the sulfate concentration values. The active mine values are consistently greater than those of the orphan mine area. This is interpreted as the result of the oxidation of freshly exposed sulfide-bearing rocks in the mining process. Hence, an increase in sulfate is a rather good indicator of land disturbances where there are rock bearing sulfides. Average sulfate concentrations caused by mining were 1705% over background at PC-2 and 6561% over background at AM-2. The residual effects downstream on the basis of average sulfate concentrations were 562% above

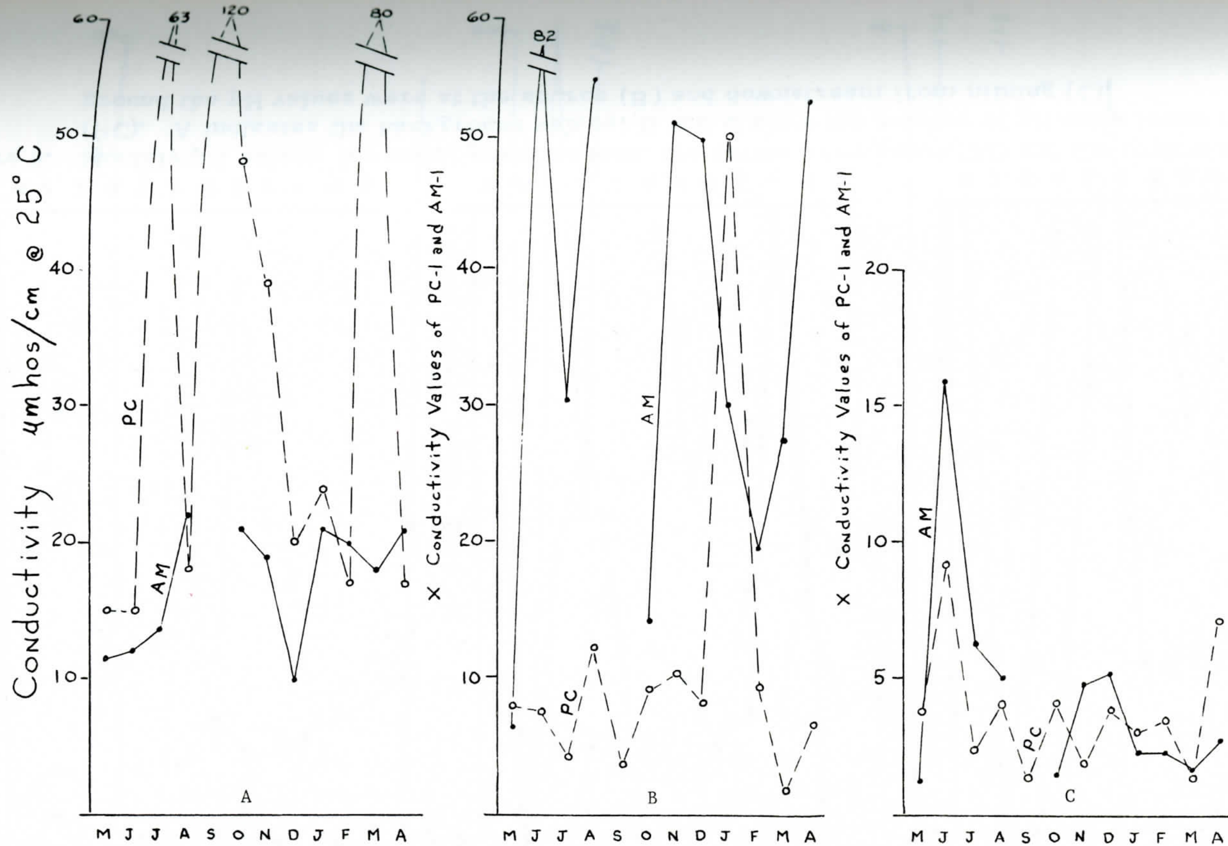

Figure 4. Monthly conductivity values for water samples from the active mine area (AM) and the orphan mine area (PC). A indicates the background values; B and C show the number of times greater than background the conductivity values were at the source (B) and downstream from mining (C).

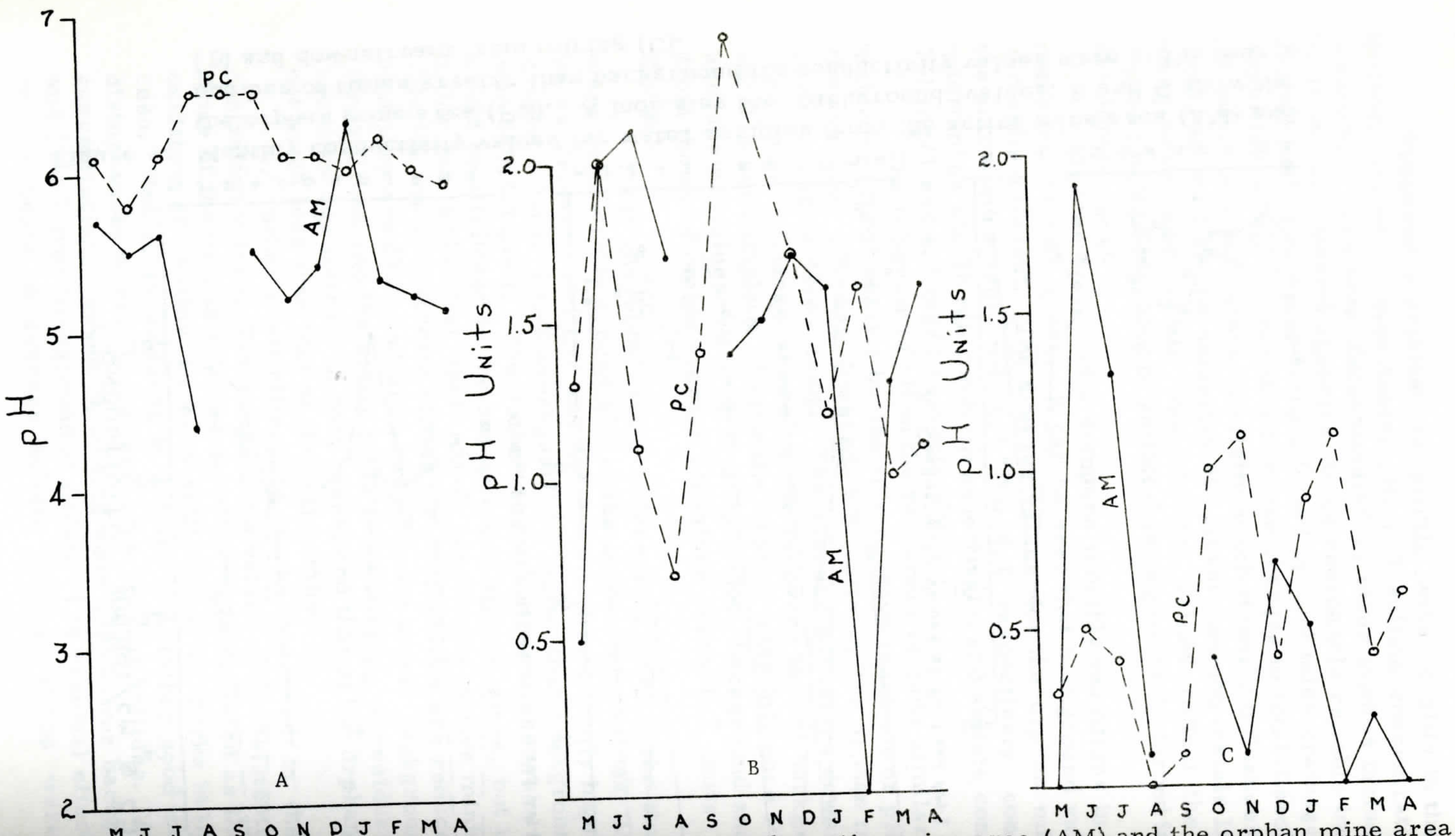

Figure 5. Monthly pH values for water samples from the active mine area (AM) and the orphan mine area (PC). A indicates the background values; B and C show the number of pH units lower than background the pH values were at the source (B) and downstream from mining (C).

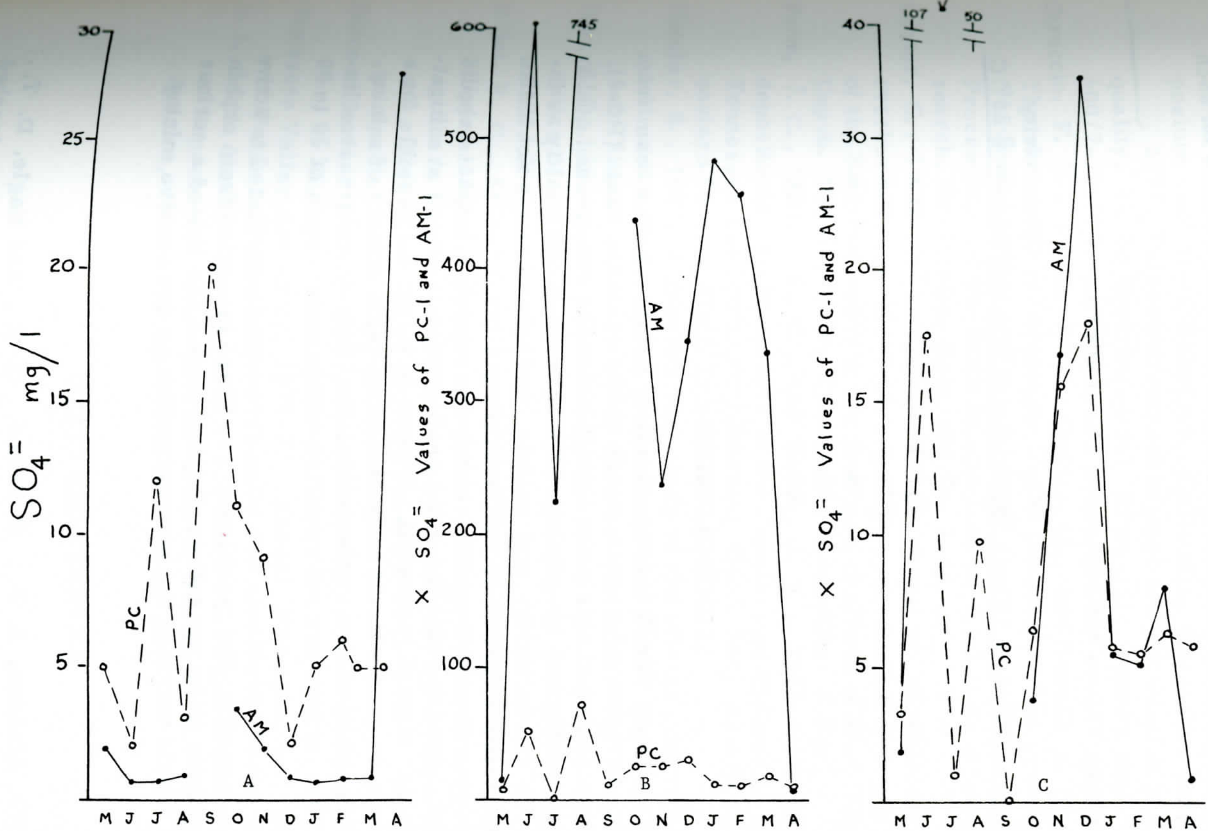

Figure 6. Monthly sulfate concentration values for water samples from the active mine area (AM) and the orphan mine area (PC). A indicates background sulfate values; B and C show the number of times greater than background the concentrations were at the source (B) and downstream from mining (C).

background on Piney Creek and 663% above background in the active mine stream.

Table 1 is a summary of the average values for the parameters measured for the monthly samples collected at the three points on each of the two areas.

Table 1. Average Values for Measured Parameters

Sample Pt.	pH	$SO_4^{=}$ ppm	Conductivity umhos/cm @ 25° C
PC-1	6.1	7.1	40
PC-2	4.3	121.1	317
PC-3	5.3	39.9	108
AM-1	5.1	4.1	16
AM-2	3.5	269.0	607
AM-3	4.4	27.2	82

CONCLUSIONS

PC-3 on Piney Creek is approximately 9 miles downstream from the major concentration of orphan mines in the drainage basin (PC-2), yet the quality of water at this point is well below the background quality of the drainage unaffected by mining. The change in the quality as inferred from pH, conductivity, and sulfate concentrations is not unlike the change in quality inflicted by active mining.

In the case of active mining, however, expected impacts should be short-term, because enforced mine reclamation is aimed at mitigating mine-related stream pollution. Where it might be possible for a riparian system to purge itself from the short-term effects of mining, especially when abetted by reclamation, the long-term stresses imposed by orphan mines are not rectified within even a span of 20 to 30 years.

Water quality problems are often erroneously attributed to active mining, when actually the problems are those inherited from orphan mines in proximity to an operation. Orphan mines are indeed a serious problem and should warrant the same attention as does active mining.

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THE MISSISSIPPIAN-PENNSYLVANIAN SYSTEMIC

BOUNDARY IN EASTERN KENTUCKY

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ABSTRACT

Controversy still exists concerning the nature of the Mississippian-Pennsylvanian systemic boundary in eastern Kentucky. Until recently, most workers had accepted or assumed the existence of a major time break across the boundary; however, recent sedimentological and structural investigations have revealed areas where the presence of an unconformity is very doubtful. Mississippian-Pennsylvanian rocks are exposed along the northeast-southwest trending "Pottsville escarpment: from the vicinity of Portsmouth, Ohio, to Burkesville, Kentucky, near the Tennessee border. The Kentucky River fault system and the Irvine-Paint Creek fault system are two northeast-trending structural features which affected erosional and depositional environments in eastern Kentucky during the Carboniferous. In the vicinity of these structural zones paleokarst, thinning and thickening of stratigraphic units and paleochannels clearly indicate the presence of an unconformity. Away from the structures the "systemic boundary" appears to be transitional, with marine units of the Mississippian grading upward into nonmarine and marine strata of Pennsylvanian age. Presently no reliable biostratigraphic data have been offered to aid in the resolution of the boundary problem. Such information will be difficult to obtain because the Mississippian rocks are primarily marine, whereas the Pennsylvanian rocks are predominantly nonmarine. Physical evidence indicates that the nature of the Mississippian-Pennsylvanian boundary is related to structure and that no regional unconformity marks this boundary in eastern Kentucky.

INTRODUCTION

Campbell (1898a) in the description of the Richmond folio described an unconformable relationship between the Mississippian and Pennsylvanian Systems. This relationship is based upon the observation of erosional surfaces between Mississippian and Pennsylvanian rocks and the lack of plant fossils indicative of Lower Pottsville rocks (Campbell, 1898). Also, without exception, United States Geological Survey maps of eastern Kentucky through 1977 show an unconformity; therefore, the existence of an unconformity between the Mississippian and Pennsylvanian Systems has been generally accepted not just for the area described by Campbell but for the entire western area of the Appalachian Basin.

The presence of a widespread unconformity has recently been questioned based upon detailed sedimentological and structural investigations (Horne and Ferm, 1970; Haney and others, 1975). The Mississippian-Pennsylvanian systemic boundary obviously exists somewhere within the geologic section of eastern Kentucky; however, since its nature has been recently questioned,the basic question exists--is the boundary unconformable or is it a conformable transition from marine Mississippian to nonmarine Pennsylvanian rocks? The purpose of this paper is to attempt to reveal the actual nature of the boundary by showing that in certain outcrop areas of eastern Kentucky the stratigraphic sequences are unconformable whereas in other areas they are transitional.

Where the boundary is unconformable, it appears to be directly related to a series of east-northeast trending fault zones which have caused a series of structural highs and adjacent lows. These structures were active during Mississippian and Pennsylvanian time, and features characteristic of unconformities, such as paleochannels, paleosinks, collapse breccias, and formational truncations, developed on the structural highs. In the adjacent low areas, no such features occur and the systematic boundary is transitional.

Features characteristic of unconformities are well displayed in Menifee County, Kentucky, in the vicinity of the Kentucky River fault system (Woodward fault zone of Silbermann, 1971), as well as along the Irvine-Paint Creek fault to the south (Figure 1). Features indicative of tectonic activity along the "Woodward fault zone" during the time of Mississippian and Pennsylvanian deposition have been reported by a number of workers. Dever (1973) and Dever and MacQuown (1974) observed erosional thinning or complete removal of lower Newman and upper Borden units on the upthrown fault block (Figure 2). Ettensohn (1974) reported facies variations in the upper Newman which he attributed to tectonic activity along the fault. Thickening of Pennsylvanian clastic units was recognized on the south (downthrown) side of the fault, and paleokarst, paleochannels, and collapse breccias were noted in the vicinity of the fault (Haney and others, 1975). Southward from the vicinity of the Kentucky River and Irvine-Paint Creek fault systems these features are not present. Hoge (1976) reported the occurrence of chert in the Newman Limestone in the vicinity of the Kentucky River fault system in Menifee and Rowan Counties (Figure 1) which he considers related to Mississippian tectonic activity. Dillenberger (1976) noticed facies variations in the Haney and Glen Dean units (upper

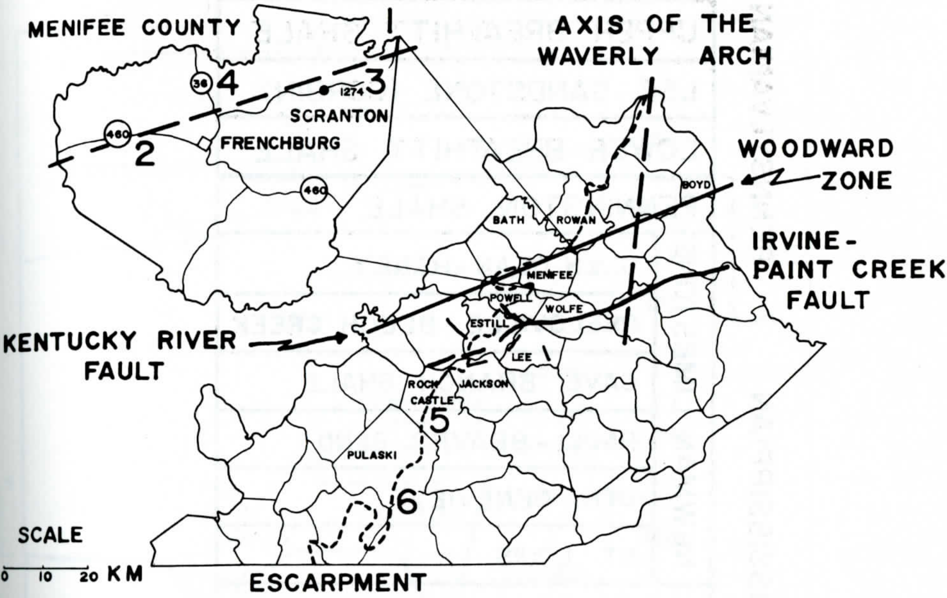

Figure 1. Sketch map of eastern Kentucky showing major structural features and locations of selected stratigraphic sections across the Mississippian-Pennsylvanian boundary. Numbers 1 through 6 correspond to the sections illustrated on Figure 3.

Newman) in the same area which he attributed to tectonism and which support observations by earlier geologists (Butts, 1922; Weller, 1931; McFarlan, 1943).

Acknowledgments

I wish to acknowledge Garland R. Dever, Jr. and John C. Philley for their constructive criticism of the manuscript. Also, I thank Norman C. Hester, Harry P. Hoge, Charles T. Helfrich, and Donald W. Hutcheson for their comments and criticisms.

DISCUSSION

General Stratigraphy and Structure

Well-defined sequences of Upper Mississippian and Lower Pennsylvanian rocks crop out in eastern Kentucky as relatively narrow zones along the Pottsville escarpment from the vicinity of Portsmouth, Ohio, on the Ohio River to Burkesville, Kentucky, near the Tennessee border

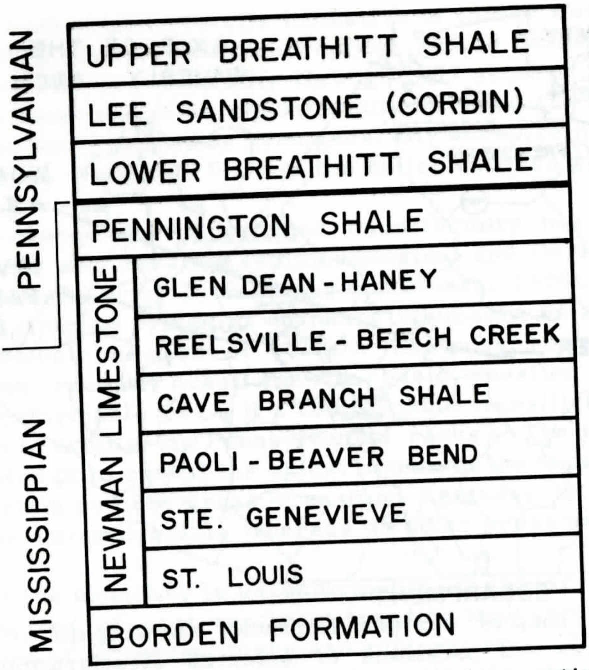

Figure 2. Generalized stratigraphic section of the Upper Mississippian and Lower Pennsylvanian rock sequences of eastern Kentucky.

(Figure 1). Mississippian sequences in ascending order include marine shale and dolomite of the Borden Formation and marine limestones and shales of the Newman and Pennington formations. Pennsylvanian rocks include sandstones, siltstones, and shales of the Breathitt and Lee Formations which are predominantly nonmarine. Differentiation between Mississippian and Pennsylvanian rocks in eastern Kentucky has historically been based upon presence or absence of marine and nonmarine fossils. The rock sequences have a slight southeasterly regional dip off the Cincinnati Arch into the Eastern Kentucky coal field (Appalachian Basin).

The Kentucky River fault system, or Woodward fault, is a major structural feature in the area of study (Woodward, 1961, Figure 1). Existence of this fault in the subsurface is well documented (Thomas, 1960; Woodward, 1961; Webb, 1969); however, surface offset is not evident. Pre-Mississippian faulting in this area was reported by Thomas (1960), Webb (1969), Silberman (1971), and Black and Haney (1975). Dever (1973) and Ettensohn (1975), after detailed stratigraphic studies of the Newman Limestone, reported evidence of structural influence of Late Mississippian deposition. Haney and others (1975) and

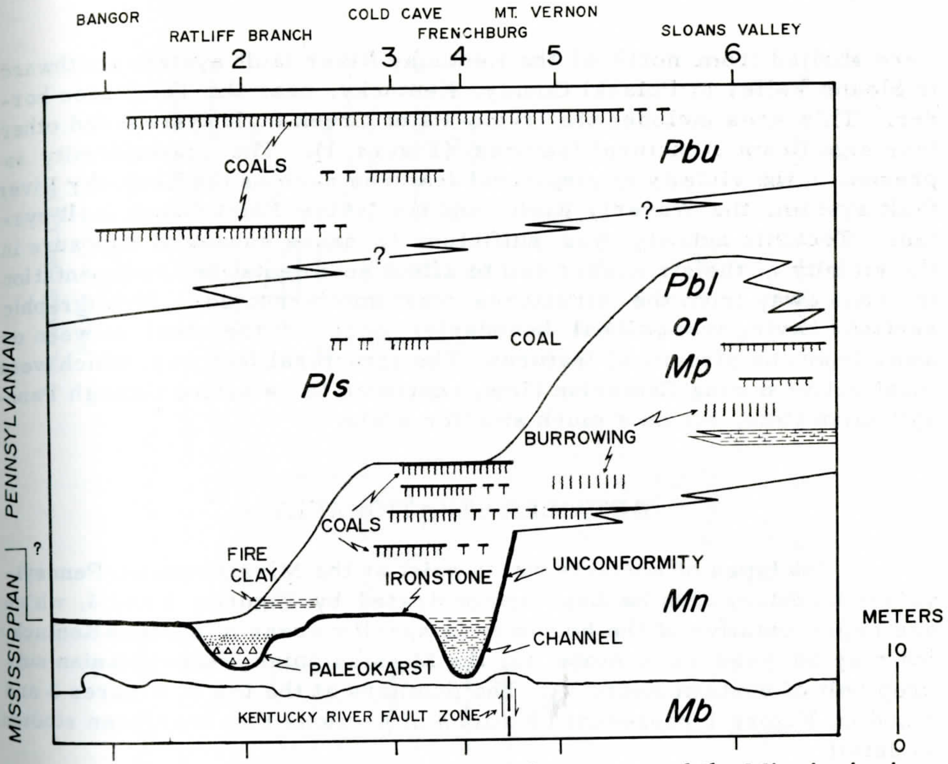

Figure 3. Diagrammatic illustration of the nature of the Mississippian-Pennsylvanian boundary in eastern Kentucky. Numbers 1 through 6 represent various types of boundary conditions observed in eastern Kentucky. Mb, Borden Formation; Mn, Newman Limestone; Mp, Pennington Shale; Pbl, Lower Tongue of the Breathitt Formation; Pls, Lee Sandstone; Pbu, Upper Member of the Breathitt Formation.

Hester and others (1975), during detailed quadrangle mapping in the outcrop area, recognized features characteristic of subaerial exposure in the Mississippian Newman Limestone along the Woodward zone. Also, they noted thickening of the lower Breathitt shale southward from the structurally active area, which further documents Late Mississippian-Early Pennsylvanian tectonic activity.

Detailed studies of the Mississippian-Pennsylvanian section in other areas of eastern Kentucky also reveal features characteristic of subaerial exposure such as paleochannels, erosional truncation, and paleokarst. Observations also reveal stratigraphic sections that are transitional from marine to nonmarine across the so-called Mississippian-Pennsylvanian unconformity with no obvious evidence of an erosional break (Figure 3 and Figure 4, columns 5 and 6). Outcrops were studied from north of the Kentucky River fault system southward to Sloans Valley in Pulaski County, Kentucky, near the Tennessee border. This area includes the Irvine-Paint Creek fault system and other less significant structural features (Figure 1). The unconformity is present in the vicinity of structural features such as the Kentucky River fault system, the Waverly arch, and the Irvine-Paint Creek fault system. Tectonic activity was sufficient to cause subaerial exposure in the vicinity of the structures and to affect sedimentation; sedimentation in areas away from the structures was uninterrupted. Stratigraphic sections having transitional boundaries occur in the areas between or away from the structural features. The structural features, which were most active during Cambrian time, continued to be active through Pennsylvanian time, but on a much smaller scale.

DETAILED STRATIGRAPHY

The types of contacts which exist at the Mississippian-Pennsylvanian boundary can be best demonstrated by Figures 3 and 4, which are representative of the boundary in specific areas of eastern Kentucky but may be used as a model for the Mississippian-Pennsylvanian outcrop belt of eastern Kentucky. The numbers at the top of Figures 3 and 4 and on Figure 1 represent locations of sections that have been studied in detail.

Section 1, is an example where sandstone of the Lee Formation (Corbin Sandstone) rests directly on the Newman Limestone; the lower Breathitt shale is absent.

Section 2 represents a contact between a karst surface of the Upper Mississippian Newman Limestone and the Lower Pennsylvanian lower Breathitt shale. The paleosinks, which contain block chert breccia from the St. Louis Limestone, generally occur low in the Newman Limestone and are overlain by sandstones and shales of the lower Breathitt.

Section 3 illustrates a boundary along which ironstone deposits developed on the subaerially exposed limestone prior to deposition of Pennsylvanian clastic sediments. Similar ironstone deposits occur along other areas of the Kentucky River fault system and the Irvine-Paint Creek Fault system in Estill and Lee Counties, approximately 20 miles to the south (Figure 1).

Section 4 shows a Pennsylvanian paleochannel cut deeply into the underlying Mississippian limestone. Sediments in the channel grade upward from sandstones into dark shales of the lower Breathitt.

The sections just described clearly indicate the existence of an unconformity.

Sections 5 and 6 represent continuous sequences reflecting a transition upward from marine to nonmarine environment in areas away from fault zones. Other geologic sections representing continuous

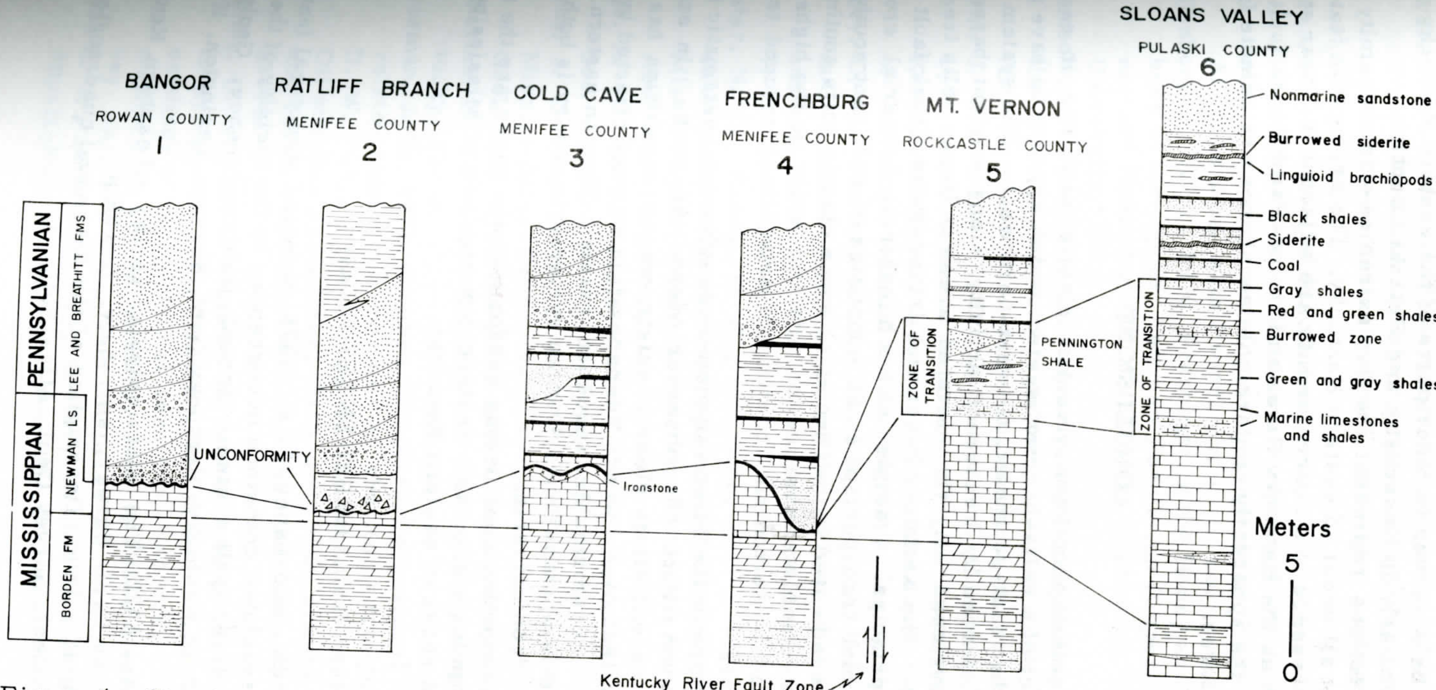

Figure 4. Stratigraphic sections of Mississippian-Pennsylvanian sequences of selected areas in eastern Kentucky. Numbers and locations at the top of each section correspond to number and location designations in Figure 3.

deposition may be observed in other parts of the eastern Kentucky outcrop belt, particularly in Rockcastle and Pulaski Counties.

These sections represent the two extremes--unconformity and conformity--yet all occur in eastern Kentucky. The areas with features indicating the presence of an unconformity are adjacent to linear structure zones such as the Kentucky River and Irvine-Paint Creek fault systems, whereas the apparently conformable sequences are remote from these same features.

CONCLUSIONS

The Mississippian-Pennsylvanian clastic sequence in eastern Kentucky is part of a delta-barrier system which appears to have prograded from the east-southeast to west-northwest. The system was occasionally disturbed by recurrent activity along ancient basement faults which developed slightly elevated areas in linear belts trending east-northeast. The Kentucky River and Irvine-Paint Creek fault systems are the principal examples of such features. Subaerial erosion of these highs and transport of small quantities of sediment southward off the highs is indicated by thickening of some clastic units southward and thinning or complete absence of specific units on the highs. In areas unaffected by tectonism, continuous deposition produced transitional boundaries.

The nature of the Mississippian-Pennsylvanian systematic boundary depends upon where the observer views the rocks. In eastern Kentucky, an unconformity clearly exists where deposition has been interrupted by tectonism; where the deposition was not affected by tectonic activity, a conformable sequence exists. Thus in eastern Kentucky, the Mississippian-Pennsylvanian systemic boundary is both conformable and unconformable.

Data supporting conclusions set forth in the paper are the result of detailed mapping within the western outcrop belt of Mississippian-Pennsylvanian rocks of eastern Kentucky.

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THE CHARACTERISTICS AND ORIGIN OF THRUST SLICES

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ABSTRACT

A slice is a block of rocks caught along a thrust fault where the strata comprising the slice are intermediate in age between the hanging wall rocks and the footwall rocks.

Slices occur along every major fault in East Tennessee. The study area was chosen because of the number of previously reported slices and the fact that many of these have been mapped in great detail.

Detailed investigations of numerous slices reveal similar characteristics for all slices relative to their distribution, orientation, mode of origin, and lithology. Slices are generally composed of competent carbonate rocks, overturned, intensely fractured, and most commonly originate from the footwall of the thrust.

Dimensions average 1. 6 kilometers, by 200 meters with a maximum known length of eleven kilometers.

INTRODUCTION

Thrust faults are the dominant structural features in the Valley and Ridge Province from southwestern Virginia to north Georgia. Although volumes have been written on the controversial subject of the origin of thrusts (see for example Rodgers, 1953a, Gwinn, 1964), the purpose of this paper is not to critique those ideas but to present my own. The present outcrop pattern of the region indicates that these large thrusts originated in linear overturned anticlinal folds which broke subparallel to the axial plane (Figure 1). Thrust slices and secondary thrusts may form parallel to the axial surfaces of overturned synclines. Strata are commonly overturned in slices. This paper deals primarily with the characteristics and source of thrust slices.

DEFINITION OF THRUST SLICES

A search of the geologic literature reveals a lack of papers which deal primarily with the topic of slices, and, moreover, most

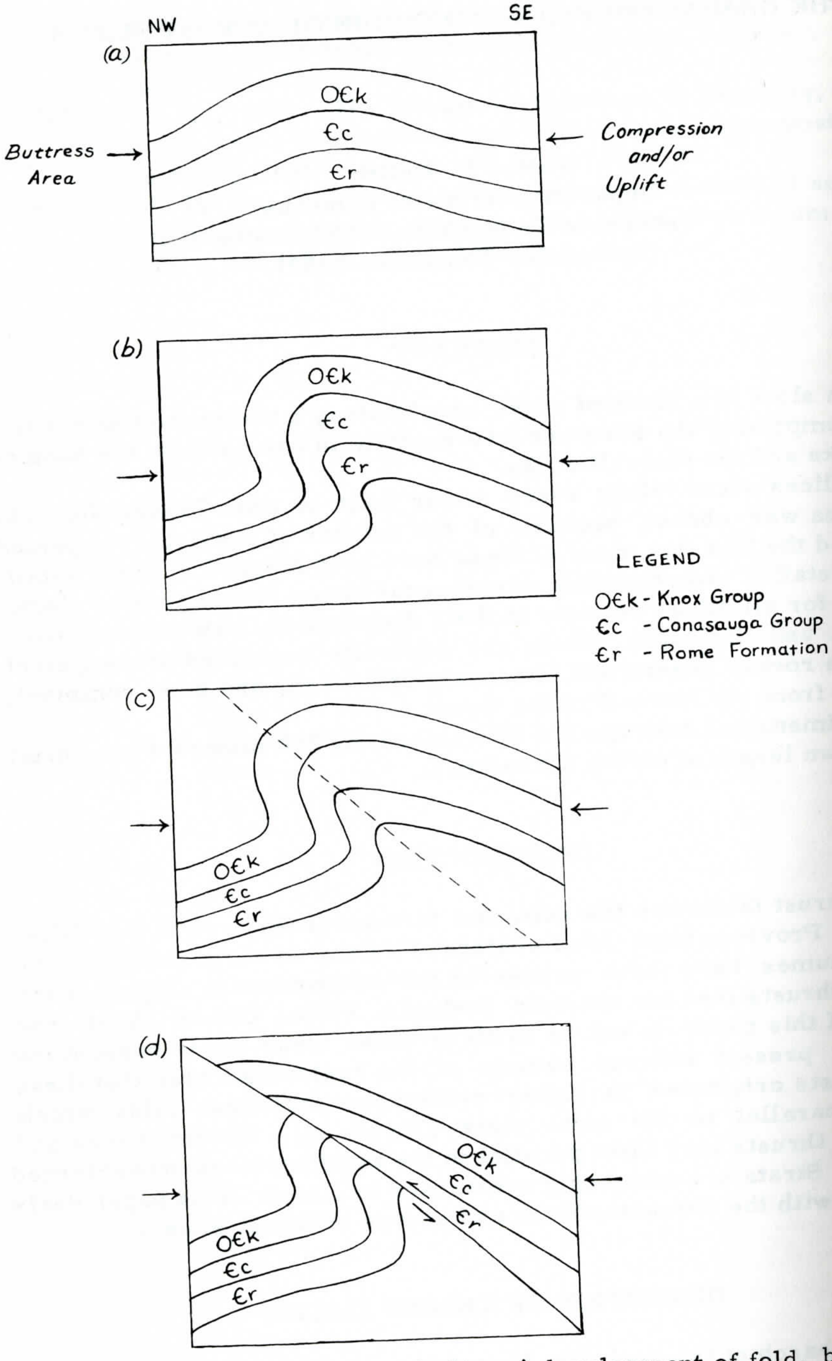


Figure 1. Origin of thrust faults; a) development of fold, b) formation of asymmetric folds, c) failure occurs along axial plane, d) thrust movement.

papers dealing with orogenic belts which contain recognizable slices have not treated the subject as a phenomenon related to thrusting. The term slice has been used to denote various entities by different writers (Hake and others, 1942; Rodgers, 1953a; Hume, 1957; Milici, 1962; Haney, 1966; and Harris, 1970). The term has often been used when referring to features in an area where successive subparallel thrusts have occurred, for which the terms thrust block, thrust belt, thrust sheet or thrust plate, all of which are more indicative of large scale faulting, are more appropriate.

The Glossary of Geology (1972) implies that the usage of the term slice should be synonymous with nappe. However, the Dictionary of Geological Terms (1962) states that a slice is a large block caught along a thrust. Billings (1954) also uses the term slice for blocks caught along thrust faults. This writer thinks that the term slice should be restricted to the latter usage with one additional restriction, the strata comprising the slice must be intermediate in age between the hanging wall rocks and the footwall rocks.

GENERAL CHARACTERISTICS OF FAULT SLICES

Listed below are features commonly displayed by slices occurring along the major overthrusts in the Valley and Ridge Province of Tennessee.

1. Slices are generally wedge or football shaped.
2. The hanging wall thrust plane has a steeper dip than the footwall thrust plane.
3. Rocks within the slice are intermediate in age between the overlying allochthonous rocks and underlying rocks.
4. Structure of the footwall side is commonly synclinal.
5. Strike of the main fault and strike of the slice are parallel to subparallel.
6. Slices occur in groups along a fault.
7. Commonly, the hanging wall rocks are fine-grained clastics.
8. The upper thrust plane is commonly warped in the vicinity of the slice.
9. Deformation is more intense close to the slice and in most cases the slice is highly contorted and overturned.
10. Slices of the Knox Group are more abundant than other units.
11. The greatest concentration of slices occur along the White Oak Mountain fault.
12. Slices average about 1.6 kilometers in length with a length to width ratio of 8 to 1.
13. Slices are composed of competent rocks, mostly carbonates.

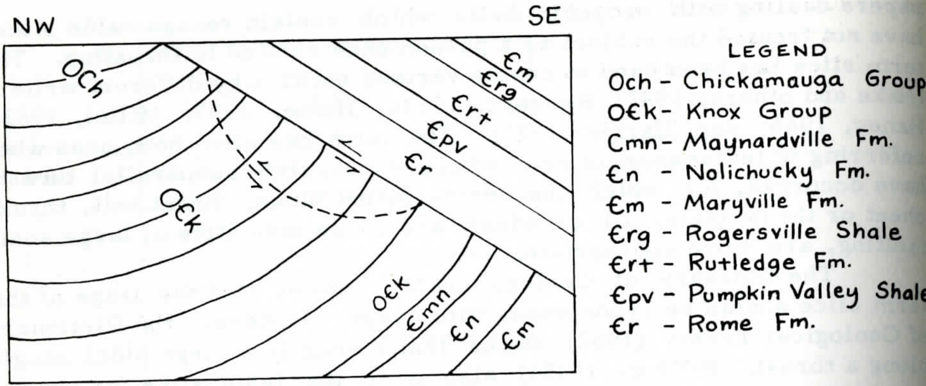


Figure 2. Schematic cross-section showing source area of slice of synclinal footwall of competent units.

ORIGIN OF SLICES

One of Dr. Swingle's favorite classroom expressions relative to the origin of thrust sheets in the Valley and Ridge was that. . . "Thrusts are developed from broken asymmetric anticlines. After variant movement and subsequent erosion, the hanging wall is viewed as a bedding plane fault." Fleck (1970) states that the most favored mechanism involves the development of thrust faults from strongly asymmetric folds or recumbent nappes. However, it appears that the alternating competent to incompetent lithologies present in the Valley and Ridge in Tennessee preclude thrusts developing from recumbent folds.

The writer has concluded from the study of more than 100 cross sections that slices originate from the footwall of thrust faults through brittle deformation and fracturing of competent rocks by pressure from the hanging wall (Figure 2). However, hanging-wall-derived slices would allow for some imaginative interpretations and might possibly explain the origin of some reentrants.

Despite the incomplete understanding of the origin of slices, there is sufficient information from which to ask logical questions and to draw some tentative conclusions. Does the evidence suggest that slices are more abundant along faults where the fault plane has refracted upward across a competent unit? Are slices more common along certain faults because the present level of erosion is at the ramp level? Is the slice overturned during movement or is it derived from an overturned limb of a syncline? Do the fault planes surrounding a slice dip more steeply with the erosion level at or slightly above the ramp level? Would a slice which has been taken from the top part of the ramped footwall show gentle dips if exposed along a higher glide plane? A companion question possibly of greater importance concerns the vacated area from which the slice was derived. What anomalous features would be observed if erosion progressed sufficiently to expose the zone from which the slice was removed? How would the outcrop pattern differ from adjacent areas along strike? Would the hanging wall rocks collapse or fold downward and occupy the area vacated by the slice? If so, how would the geometry of the overlying fault block be affected? Would the local strike of the fault be changed significantly, creating a salient and/or a rerentrant?

EXAMPLES OF SLICES IN TENNESSEE

General Discussion

The Geologic Map of East Tennessee (Rodger, 1953b) depicts more than 50 easily recognizable slices as defined by this writer. Although all of the major thrust faults (Figure 3) contain slices, the White Oak Mountain fault contains 16 slices which represent almost one-third of the total slices shown. Other major faults which contain abnormally large numbers of slices are the Dumplin Valley, Saltville, and Chattanooga faults.

Practically every slice is composed of competent rocks, mostly carbonates. Rocks of the Knox Group are present in about 40 of the 50 slices and most of the other slices are either the Honaker Dolomite or limestones of the Chickamauga Group. Clastics of the Rome Formation form the hanging wall above some 30 of the slices and many others have fine clastics of the Conasauga Group in the hanging wall. The dip of the footwall beds beneath most slices is less than 10^{o}, whereas the dip of the hanging wall beds over the slices is 30^{o} or more. The strike length of slices may be only a few hundred meters or several kilometers in length. However, dimensions of surface exposures show that slices average 1.6 kilometers along strike and 200 meters in outcrop width, a ratio of 8 to 1. Byerly (1966) describes a slice along the Pulaski fault which is 600 meters wide and over eleven kilometers in length.

Characteristics of Selected Slices

Geologic quadrangle maps of East Tennessee give a more detailed picture of the characteristics of slices. Cross-sections accompanying these quadrangles appear subjective in that some geologists utilize slices while others do not in projecting the same pattern of surface geology. Moreover, if one looks only at cross-sections instead of geologic maps, small imbricate or divergent faults may easily be mistaken for slices. However, by definition of a slice, it is necessary that the secondary fault intersect the main fault trace.

The Roddy Quadrangle (Milici and Swingle, 1972) shows a large area between the Rockwood and Chattanooga faults, the cross-section of which exhibits many of the classical features of a slice (Figure 4).

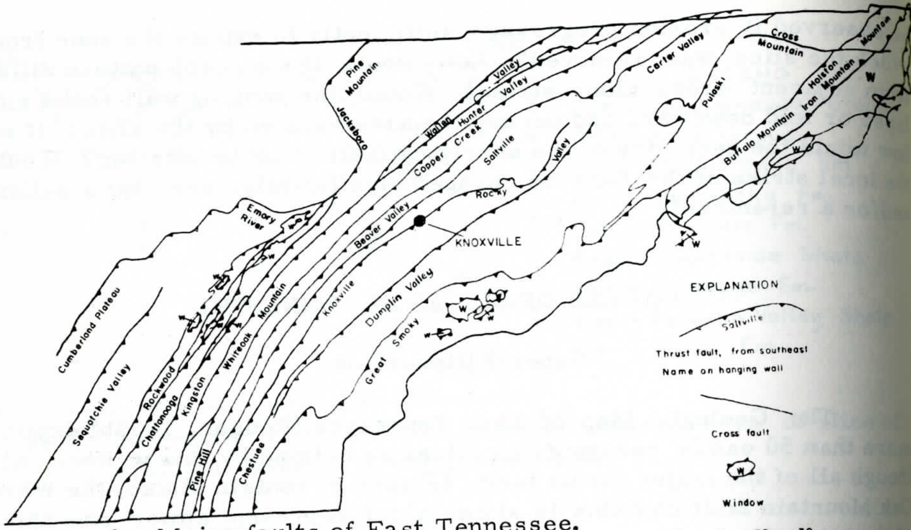


Figure 3. Major faults of East Tennessee.

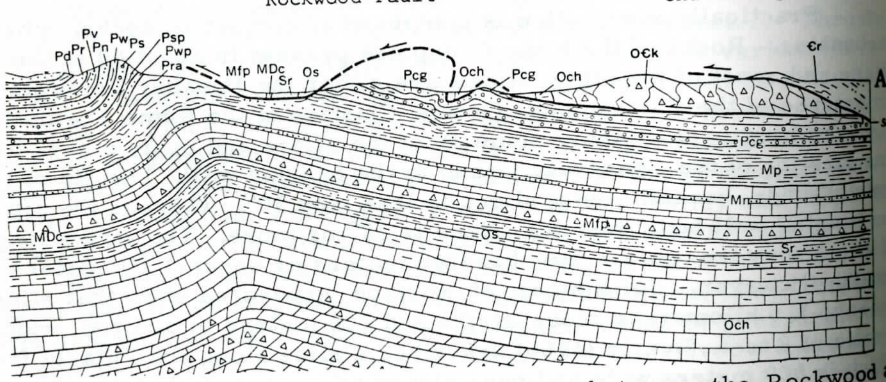


Figure 4. Geometry of a "classical slice between the Rockwood and Chattanooga faults (after Milici and Swingle, 1972).

However, without viewing the contiguous geologic maps, one might simply interpret the structure of this area as imbricate or piggyback thrusting. Other similar areas of possible slices can be seen on Morgan Springs 7 1/2' Quadrangle (Swingle, 1963), Evensville 7 1/2' Quadrangle (Swingle, 1964a), Clinton 7 1/2' Quadrangle (Swingle, 1964b), Spring City 7 1/2' Quadrangle (Swingle, 1964c), Graysville 7 1/2' Quadrangle (Swingle, 1964d), and other geologic quadrangles which border the Eastern Cumberland Plateau Escarpment.

Smith (1976) suggested that some structural features mapped as klippen northwest of the Chestuee fault in McMinn County, Tennessee could also possibly be interpreted as vestigial "slices" from which the

A. Thrust plane relatively undistorted. Compressional forces active

B. Compressional forces continue causing folding of the fault plane until it locks the fault at which time a second break occurs along the dashed line forming a slice.

C. Erosion subsequent to folding-faulting removes the allochthonous block and separates and exposes the slice which then becomes a klippe.

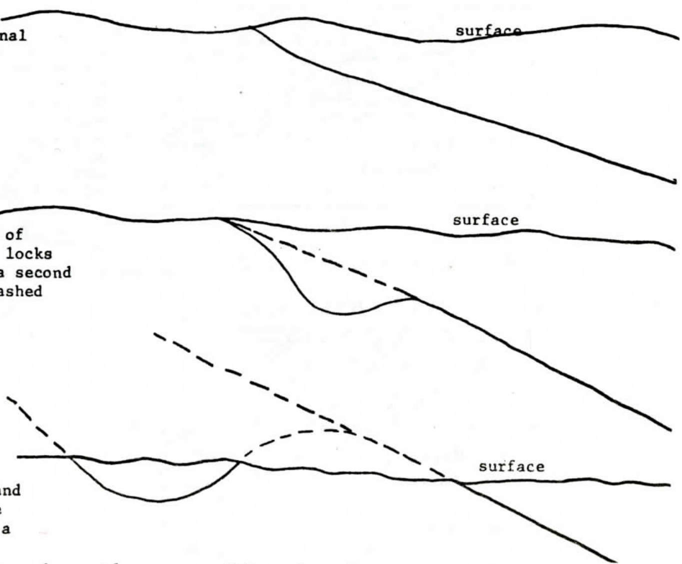
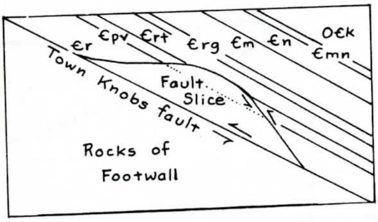


Figure 5. Sketch showing the possible development of klippen from slices (after Smith, 1976).

overlying fault and encompassing rocks have been eroded. To class these structures as klippen, requires a relatively sharp dip reversal of the fault plane from southeast to northwest, so that the fault plane passes beneath these structures. Smith suggested that the "slices" broke from the footwall of the main thrust fault during folding and thrusting, moved only a short distance and were overridden as movement continued (Figure 5). Penecontemporaneous folding and faulting readily explain the relationship between the relatively steep dip of the near-surface portion of the Chestuee fault and the nearly flat attitude of the fault plane beneath the "slices". Indirect supportive evidence for a slice-origin interpretation for these structures is that the beds in the klippen (slices) have strikes closely aligned to the unfaulted beds upon which they lie, a somewhat unlikely situation if the klippen were derived from the hanging wall or if transported even greater distances.

Erosion exposes fault slices at the surface. Therefore, continued erosion should also expose the area from which a slice was derived. Haney (1966) suggested that reentrants could be formed by erosion on the hanging wall if erosion exposed the vacated area (Figure 6). An alternative idea for deriving a hanging wall slice is given in Figure 7. A slice derived from the hanging wall would account for the gentle dip of the hanging wall presently exposed over the vacated area. In this case, the fault trace is eroded and retreats faster than in the areas adjacent to the reentrant. A hanging wall source for the slice would also account for the absence of the older units on the hanging wall in the reentrant area. In contrast to suspected hanging wall-derived slices,

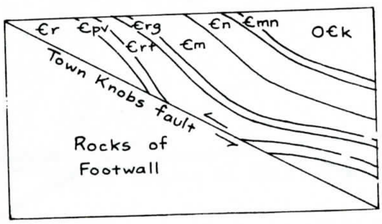


LEGEND

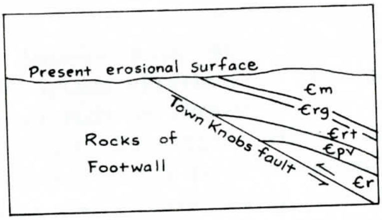
O€k - Knox Group
€mn - Maynardville Formation
€n - Nolichucky Formation
€m - Maryville Formation
€rg - Rogersville Shale
€rt - Rutledge Formation
€pv - Pumpkin Valley Shale
€r - Rome Formation

A

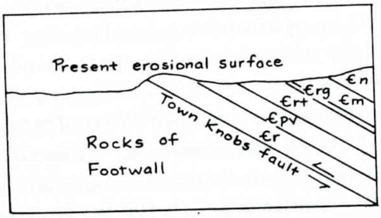
Stage 1. Fault slice breaks from the hanging wall of the fault.



Stage 2. Hanging wall moves above and beyond the slice block, thus the Rutledge Formation of the hanging wall lies on rocks of the footwall.



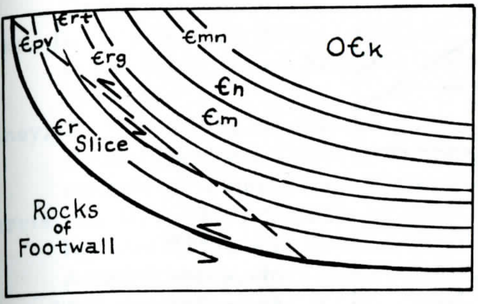
Stage 3. Present erosional surface across the reentrant shows the Rutledge Formation against the footwall of the fault.



B

Present erosional surface northeast and southwest of the reentrant shows the Rome Formation against the footwall of the fault.

Figure 6. Cross-section sketches: A. Inferred stages of the possible development of the small reentrants in the trace of the Town Knobs fault, B. Cross section through the area either northeast or southwest of the reentrants (after Haney, 1966).



LEGEND

OЄk - Knox Group
Cmn - Maynardville Fm.
Єn - Nolichucky Fm.
Cm - Maryville Fm.
Єrg - Rogersville Shale
Єrt - Rutledge Fm.
Cpv - Pumpkin Valley Sh.
Єr - Rome Fm.

Figure 7. Cross-section showing a slice derived from the hanging wall of a thrust fault.

the vacated area of a footwall derived slice would form a salient (Figure 8).

Economic Aspect of Slices

Possibilities exist that thrust slices of the Knox Group may be favorable ground in the search for zinc or barite as mineralized host rocks which have been tectonically transported away from a major producing district. Perhaps the larger slices immediately northwest of the producing districts warrant a reconnaissance drill hole program. Hydrocarbons and ground water could be contained within the fracture-generated porosity. Carbonates within slices may be suitable for aggregate or road metal.

SUMMARY AND CONCLUSIONS

1. Most recognized slices have many of the same characteristics.
2. Dimensions of surface exposures show that slices average 1.6 kilometers along strike and 200 meters in width.
3. Slices are composed of competent rocks, mostly carbonates.
4. White Oak Mountain fault contains the most slices in East Tennessee.
5. Slices commonly originate from the footwall of thrust faults.
6. Strata are commonly overturned in slices.
7. Structural features which have been interpreted as imbricate faulting, or even some named klippen, may be slices.
8. If erosion, exposed the source (vacated) area of a slice, the fault trace would deviate according to the size of the slice removed. Salients and reentrants may be explained by whether the slice originates on the hanging wall or footwall.

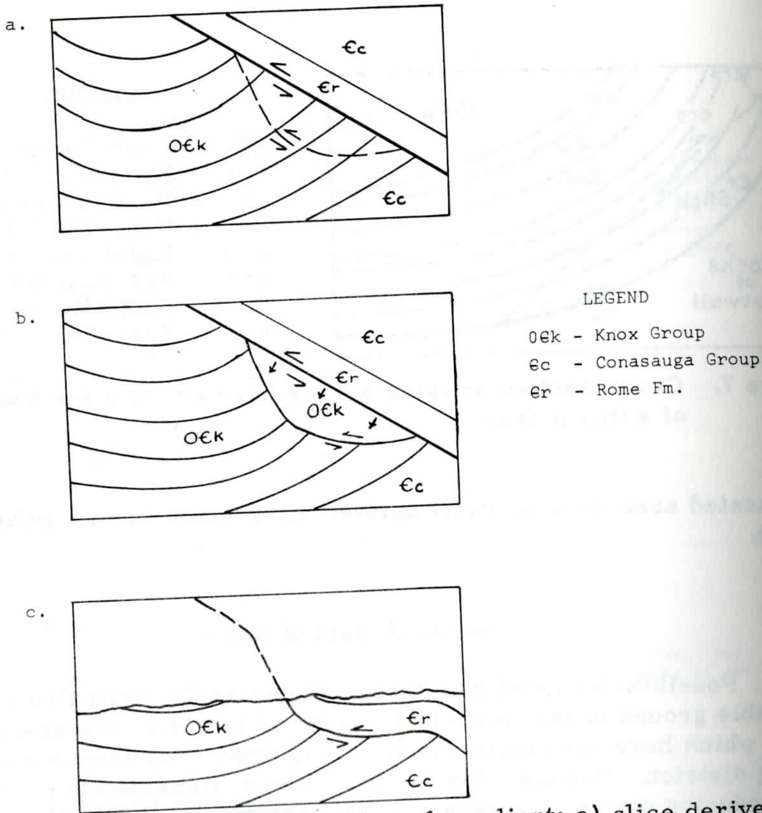


Figure 8. Possible formation of a salient; a) slice derived from footwall, b) hanging wall rocks collapse into vacated area; c) subsequent erosion and development of salient.

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GEOLOGIC AND HYDROLOGIC FACTORS FOR LOCATING REPOSITORIES FOR RADIOACTIVE WASTES[1]

By

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ABSTRACT

Rocks such as salt, shale, limestone and granite may all qualify as host media for the disposition of radioactive wastes in the proper environments. In general, the only requirement for any rock body or storage area is that it contain any emplaced wastes for so long as it takes for the radioactive materials to decay to innocuous levels. This requirement, though, is a formidable one as some of the wastes will remain active for periods of hundreds of thousands of years and the physical and chemical properties of rocks that govern circulating groundwater and hence containment, may be difficult to determine and define. Because of the complexities and variabilities of geologic environments, development of strict criteria for the selection of repository sites is not feasible. Rather, only general guidance for locating repositories is offered through a discussion of the relevant geologic and hydrologic factors.

INTRODUCTION

In a broad sense there is only one criterion for the selection of a disposal site in a rock body: the geologic and hydrologic characteristics of the site must be such that no constituents of emplaced wastes could disperse into the biosphere in hazardous concentrations for the required times of containment, which may be as long as hundreds of thousands of years. The specific geologic and hydrologic factors discussed below are merely an elaboration of this general criterion. Because of the extreme complexity of many geologic environments and the interrelated and interdependent nature of rock features that govern the presence and circulation of groundwater and the lack of experience and

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comprehensive precedents in long-term geologic disposal, it is not practical to derive rigid criteria for site selection. Therefore only very general guidelines are offered. Each potential site must, therefore, be evaluated according to its own unique local and regional geologic and hydrologic setting.

STRATIGRAPHIC AND LITHOLOGIC FACTORS

Most rocks that lie within a hundred meters of the land surface contain an abundance of open fractures that are capable of transmitting water. This, coupled with the slow but relentless removal of the land surface through erosion, makes it imperative that prospective zones for radioactive waste repositories lie at depths of at least two to three hundred meters. Thus, with the exception of some areas having arid climates, most of the waste emplacement levels will be located below the water table. Because of a pronounced acceleration in the rate of plastic flow of rock salt as the pressure increases with depth, it is prudent to restrict mechanical mining operations in salt to depths of no more than about 1,500 meters. However, solution-mined openings in salt could probably be maintained for waste disposal to maximum depths of 3,000 meters. For shale, it is expected that stable openings could not be formed below 1,500 meters from the surface. Brittle rocks such as limestone, granite and basalt could undoubtedly be mined at depths considerably greater than 1,500 meters, but the increased costs of testing, evaluating and actually developing disposal space at these great depths would appear to greatly limit their potential for repository use. From the above, it is clear that the most common depths for disposal would range from about 300 to 1,500 meters, however, slightly shallower and somewhat greater depths may also be suitable under certain conditions.

In general, rocks having a high degree of homogeneity are superior to those for storage of radioactive wastes. However, the limits of acceptability of inhomogeneity cannot be stated unequivocally but can vary greatly depending on such things as the type of host rock, the nature and characteristics of the "contaminating" rock minerals, the types of waste and the design of underground waste emplacements. For sedimentary rocks like limestone, salt and anhydrite, it is common to find rather homogeneous layers that are 30 centimeters or more in thickness separated by thin partings of shale and/or other rocks or minerals. The nature and location of these partings are important in mining and frequently, when they are widely spaced, are used to accommodate a smooth and continuous roof for the workings. On the other hand, closely spaced shale partings may promote roof instability as the bulk of the rocks tends to separate along these partings and subsequently fails.

Although the primary geologic containment for radioactive waste emplacements in rock bodies exists, for the most part, within the host rocks, it is apparent that additional protection and/or containment may be gained through impervious beds that might surround the host rocks. Thick beds of shale and/or other plastic-behaving rocks that are impermeable to the circulation of groundwater are generally preferred as they may deform without fracturing to accommodate the subsidence and/or buckling, for instance, due to thermal expansion of the host rocks. Perhaps least desirable surrounding rocks are those that contain large quantities of circulating groundwater, especially in the overlying beds, as these waters have the potential for flooding the waste-filled cavities should unexpected fracturing occur in the host rocks. Most rocks contain some open passageways for circulating water near the land surface but where separated by 300 meters or more of impervious rocks, the potential hazard of this water to the emplaced wastes should be minimal.

STRUCTURAL FACTORS

In bedded sedimentary rocks, except salt diapirs, the optimum inclination of strata for waste emplacement is less than a few degrees, as the design of underground workings and subsurface waste transport operations favor low haulage grades and nearly horizontal mining levels. Also, to maintain consistency and uniformity, it is desirable to restrict excavations to specific stratigraphic horizons or beds. Steep dips indicate that the rocks have been subjected to severe tectonic stresses that commonly fracture brittle rock and create passageways for the circulation of groundwater. Thus, even though underground openings could probably be designed to accommodate steeply dipping beds, their utility for waste disposal would be lessened as a result of the increased potential for circulation of groundwater. Also, the "protective" covering of impermeable, overlying beds would not generally be present in steeply dipping formations. For plastic-behaving rocks, like salt and some argillaceous sediments, deformation accompanying moderate to steeply dipping beds would tend to have a minimal effect on water movement as fractures would be healed and thus not open to circulating groundwater. However, for brittle rocks like limestone and anhydrite, fractures would not heal readily but would enhance any movement of water in the formations.

In general, faults and joints have a deleterious effect on rocks that are candidates for waste emplacement. They can effect the initial mining as well as the structural stabilities of such openings once they are excavated. More importantly, they constitute the principal pathways for water circulation in many rocks. Major faults that can be traced for hundreds of kilometers on the earth's surface generally disrupt rocks over broad zones at least several kilometers in width and most assuredly should be avoided entirely in selecting areas for waste repositories. On the other hand, some minor flexuring occurs in most rocks and its significance in siting can only be evaluated in light of such things as its nature and extent, rock types, hydrology, etc. In general, faults and joints are more open and thus freer to transport groundwater near the land surface than at depths of hundreds of meters. However, depth alone is not sufficient to guarantee "tightness." Independent appraisals must be made of the fault and joint systems at individual sites although these perturbations adversely affect the suitability of brittle rocks for a waste repository more than the plastic behaving ones.

In extreme cases where thick bodies of salt or other natural plastic-behaving rocks exist at depth under sufficient differential pressures, mass flowage of these rocks may produce diapiric structures like domes and anticlines. These structures may be especially suitable for waste disposal, provided that it can be determined that they are stable and flowage will not be rejuvenated during the hazardous lifetime of the wastes. In general, the energy from high heat-generating wastes affects too small a volume of the rock body and persists over too short a time to initiate or reactivate diapirism. Some salt and argillaceous deposits that are thick, deeply buried and exhibit extreme surface relief require that geologic examinations be made to ascertain that incipient diapirism is not presently taking place and that geologic processes during the next few hundreds of thousands of years will not create conditions conducive to such movements.

HYDROLOGIC FACTORS

Over long periods of geologic time, surface streams may be expected to undergo radical changes in their flow regimes. Under extreme conditions of erosion, deep incisions may be cut into the rocks, while for extended periods of stream aggradation thick beds of sediments may accumulate. In any event, it is clear that the future behavior of surface streams must be predicted to ensure that geologic containment can be maintained for the required period of time.

Circulating groundwater poses the main real threat to the containment of radioactive waste placed in rocks. Thus, the nature and characteristics of water-bearing formations that lie in close proximity to potential disposal zones are critical elements in establishing the suitability of specific rock types and special study areas.

Generally, the host rock for wastes should be free of circulating groundwater although overlying and underlying formations may contain water bearing zones. For plastic-behaving rocks like salt and shale the vertical as well as the horizontal distances between circulating groundwater and waste emplacement zones can generally be less than those for brittle rocks because of the self-healing characteristics of plastic rocks. Thick bodies of rock salt that lie at relatively shallow depths commonly are in contact with circulating groundwater along their uppermost boundaries. In these instances, the disposal levels should lie at least about a hundred meters below the water zones and/or at such distances that the projected rates of dissolution of the salt will not uncover the wastes during its hazardous period. Water bearing zones are much less common at the base of thick salt formations, however, should they be present, waste emplacement zones should also lie about a hundred meters above these zones and/or at an appropriate distance to insure long term containment of the wastes.

Ideally, for hard, brittle rocks any overlying and/or underlying aquifers should be separated from the host rock by thick bodies of impermeable shales or other aquicludes. In some cases, it may be necessary to choose host rocks containing some circulating water for waste repositories. However, und r these conditions it would be imperative to demonstrate that the radionuclides, if mobilized, would move at such a slow rate that significant quantities of them would never reach the biosphere, being confined to some acceptable zone in the subsurface.

TOPOGRAPHIC CONSIDERATIONS

In general, low relief and gently sloping terrain should characterize the topography at waste repository sites. Since rail and/or highway transportation must be provided for such facilities, it is essential that topography be amenable. In some instances, the surface features of an area reflect the subsurface conditions. Thus, irregular topography may indicate complex geologic and hydrologic conditions in the subsurface while more regular surface terrain may be suggestive of flat-lying, undisturbed rocks at depth. For thick bodies of salt that may be candidates for waste emplacement, it is essential that low to moderate relief prevail at and near such sites as extreme relief may create differential loadings on the salt formation that could trigger salt flow and diapirism (Gera, 1972).

TECTONIC AND SEISMIC EFFECTS

As all rocks are adversely affected by major crustal disturbances, areas of low seismicity and tectonic stability are favored for waste disposal facilities. Surface entrances into the waste repository chambers are particularly susceptible to damage during violent earthquakes. In extreme cases, these events could lead to temporary disruption in operational activities and even possibly to breaking the geological containment for some types of rocks.

In the long term, a major risk for the integrity of a repository would be faulting through the disposal zone (Claiborne and Gera, 1974). For the more plastic-behaving rocks, such as salt and possibly clay and shale, any fractures are healed and geologic containment is sustained. However, for brittle rocks such as granite and limestone, fractures commonly lead to passageways for the free circulation of groundwater and thus to some reduction in the effectiveness of the rock to contain the emplaced wastes. In general, regions having a long geologic history of tectonic stability also exhibit a low frequency and intensity of seismic events. Conversely, those areas that have experienced mountain-building episodes in recent geologic time also are likely to have a high incidence of historical earthquake activity.

PHYSICAL AND CHEMICAL PROPERTIES

The physical and chemical properties of rocks are extremely important in establishing the utility of various types of rock bodies for the emplacement of radioactive wastes as they determine such things as the circulation of groundwater, the dissipation of radiodecay heat, the effects of radiation, and the stability of mined openings.

Rocks that possess low permeabilities are favored for radioactive waste disposal. This is primarily related to the preference for utilizing dry excavations for waste emplacement. Except for argillaceous strata and some chalks, rocks having low primary permeability also possess low primary porosities. Primary porosity (intergranular voids) is generally low (5%) for many rock types such as salt, limestone and granite while interconnecting pore space (primary permeability) generally is extremely low for these same rocks. These conditions commonly lead to dry excavations; however, should fractures exist in these rocks, their permeabilities (fracture permeability) may be increased greatly (excepting the self-healing rocks like salt and some shales) and thereby enhance the free circulation of any groundwater present.

Inclusions of small amounts of gases and liquids may occur in many evaporites. For high-level, heat generating wastes, these inclusions may affect repository operations through rock decrepitation and migration of brine toward heat sources or waste containers. For example, in some rock salt deposits at temperatures below the decrepitation point of approximately 250° C, small brine-filled cavities (size ranging from a few millimeters down to microscopic) may migrate toward heat sources (Bradshaw and Sanchez, 1969). The mechanism for cavity migration is the diffusion of sodium and chloride ions from the hot to the cold site of the cavity due to the slightly greater solubility of NaCl at higher temperatures. Migration rates are directly proportional to temperature gradient and increase with increasing temperatures. Where these cavities represent less than about 1% of the salt volume, the maximum inflow of brine around buried waste containers would be expected to be small and would probably cease after a few tens of years. Other than its effect on radiolysis and container corrosion, this moisture should be insignificant.

For most other rock types, migration of fluids by the foregoing mechanism would probably be inconsequential. However, localized fracturing of the rocks could occur should the rocks be heated beyond their decrepitation points. Isolated brine or other fluid inclusions should not adversely effect the disposal of low-heating generating wastes in rock salt or any other rock types.

Frequently, in mining in sedimentary rock sequences, small pockets of gas are encountered that cause blowouts and/or floor and ceiling buckling. In most cases, rock pressures immediately around the openings can be relieved by drilling relief holes in to the floor or ceiling. Once encountered, the gas, if it exists in small quantities, may be discharged through the normal ventilation system of the underground workings without the deleterious effects. In extreme cases, where rich beds of carbonaceous materials are interbedded with rocks like shale and salt, seeps of methane or other hazardous gases may persist which would be extremely difficult to combat in operations requiring the use of mean and equipment in the underground workings. Geophysical methods may be employed to locate gas-filled cavities prior to mining.

For all rocks that are potentially suitable for the emplacement of radioactive wastes, it must be established that the transient and permanent rock deformations (displacements, strains and stresses) induced in the rocks will not produce conditions leading to a breach of the integrity of the long-term containment. These deformations are produced by the closure of the mine openings and, in the case of high heat generating wastes, by the thermal expansion of the entire rock column (Lomenick and Bradshaw, 1969). The wastes contained in plastic-behaving rocks are eventually encapsulated by the host rock and, since open fractures do not occur, circulating groundwater is prevented from contacting the wastes.

Preferably, rocks to be utilized for the disposal of high-level, heat generating waste should possess thermal properties that promote rapid dissipation of waste decay heat and whose stabilities are not adversely affected in the presence of elevated temperatures. Rock salt possesses rather high thermal conductivities in comparison to some other earth materials but it also is known to exhibit accelerated rates of creep at elevated temperatures (Lomenick and Bradshaw, 1969b). The latter characteristic of accelerated creep is generally desirable in the case of certain plastic-behaving rocks as this advances the sealing and healing of excavated cavities that contain the wastes. Conversely, for brittle-behaving rocks, accelerated deformation at elevated temperatures would be detrimental as it might lead to fracturing and some loss of geologic containment.

Although completely dry environments are preferred for the emplacement of radioactive wastes, some water-bearing argillaceous rocks, like clay, mudstone and some shale, that transmit water only at extremely slow rates may also be acceptable host rocks. These water-bearing host rocks must possess high adsorption capacity as this property associated with the extremely low hydraulic conductivity, would assure the long-term containment of the waste. For competent rocks like limestone, granite and basalt that transmit water only along fractures and hence possess extremely low effective adsorption capacities, it is imperative that the waste remain isolated from circulating groundwater. Nevertheless, adsorption could provide a secondary geologic barrier against the dispersal of radioactivity should the primary geologic containment be accidentally or otherwise breached. In particular cases other phenomena, such as mineral precipitation and co-precipitation from groundwater might contribute to restricting the geochemical mobility of certain radionuclides. Therefore, it is important the retention potential and groundwater flow be carefully examined at each individual site.

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STRUCTURE AND TECTONICS OF THE APPALACHIAN MIOGEOSYNCLINE NEAR THE JUNCTION OF TENNESSEE, KENTUCKY, AND VIRGINIA

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ABSTRACT

Cross sections were drawn to Precambrian basement and reconstructed upward to a proposed Permian erosional level. The basis for the forms of the folds and faults is the geometry of exposed structures in outcrops, and geologic mapping data. Paleozoic undulations of the basement were a controlling influence on localization of some major structures. Others of flexural-slip origin were probably controlled by the competent Knox unit. Ramp folds occur on the northwest side of the area. Gravity most likely played a part in deforming the rocks because the basement surface dips northwestward according to this reconstruction, not southeastward as previously proposed.

INTRODUCTION

An approach to understanding the subsurface structure of the Valley and Ridge Province is needed because of the scarcity of drilling data. This paper is an attempt to utilize structural data observable at the surface in order to predict the subsurface configurations of rock units to Precambrian basement.

The location of both the cross sections extending across much of the deformed miogeosyncline and the area of detailed mapping where minor structures were observed in outcrop are shown on the index map of Figure 1. The cross sections are mostly in northeastern Tennessee but include small portions of Virginia and Kentucky. The area of detailed study was in northeastern Tennessee and included portions of Grainger, Hamblin, and Hawkins Counties. The Saltville Fault is the major structure in the area, and below it is the Greendale Syncline.

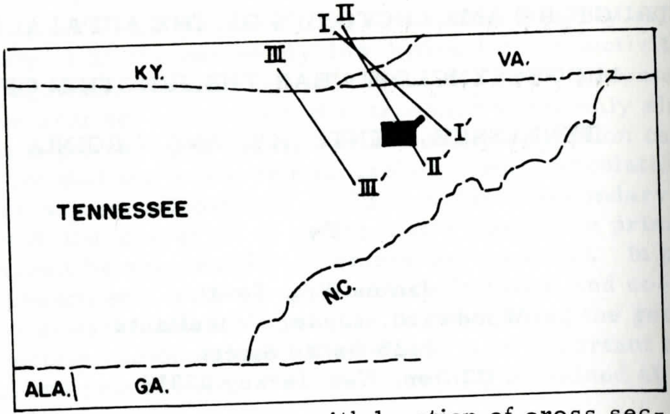

Figure 1. Index map with location of cross sections of Figure 2. Black rectangle is area of detailed mapping.

Acknowledgments

Most of this article is from a dissertation (Smith, 1968) directed by Dr. George D. Swingle, who suggested the study area because of excellent outcrops and adjacent detailed work by Sanders (1952). Dr. Swingle played a part in obtaining funds from the Tennessee Division of Geology to help pay for the field work. He accompanied the author in the field a few times and pointed out some of the major stratigraphic and structural problems. He personally furnished the author with a boat and boat trailer which proved to be a necessity in examining the outcrops along Cherokee Lake.

A very rough copy of this paper was edited by Robert D. Hatcher, Jr., Robert C. Milici and Denny N. Bearce. Hatcher also worked with the final copy. Thanks to all of them.

STRATIGRAPHY

In the area of detailed study, the exposed rocks of the upper plate of the Saltville fault range from the lower Cambrian Rome Formation through the middle Ordovician Sevier Shale. The cross sections include the Precambrian igneous and metamorphic rocks which are about one billion years old, overlain by a typical miogeosynclinal sequence of shallow water deposits and a few strata of continental origin. The rock types are mainly carbonates and shale with minor sandstone. These Cambrian through Pennsylvanian strata are about 10,000 feet thick on the west side of the miogeosyncline and are estimated to have been at least 20,000 feet thick on the east (Figure 2). The thickest structurally competent unit is in the upper part of the lower half of the

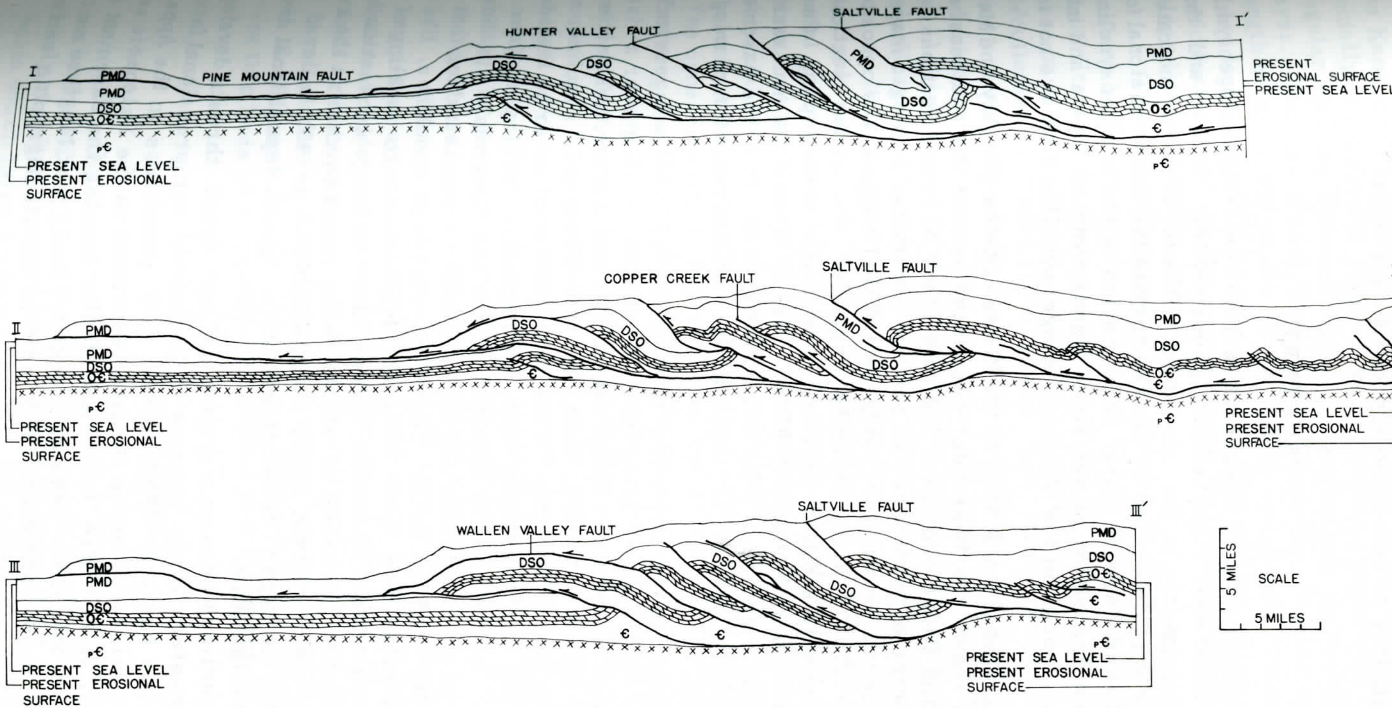

Figure 2. Permian cross sections in northeastern Tennessee and adjacent Kentucky and Virginia as shown in Figure 1. p€-Precambrian basement rocks. €-Lower to Middle Cambrian clastics. O€-Upper Cambrian-Lower Orodvician Knox Group. DSO-Middle-Upper Ordovician, Silurian and Lower Devonian rocks, undivided. PMD-Upper Devonian, Mississippian and Pennsylvanian rocks, undivided.

section, and this unit, the Knox Group, is about 4,000 feet thick, composed predominantly of dolostone and some limestone.

STRUCTURE

The study area is in the thrust faulted and folded miogeosynclinal portion of the deformed Appalachian geosyncline. Strata and thrust faults dip predominantly to the southeast. These factors indicate the miogeosyncline was deformed by nearly horizontal compressional forces from the southeast. Age of major deformation of the miogeosyncline is clearly between deposition of deformed Pennsylvanian strata and intrusion of Triassic rocks in the Virginia miogeosyncline that are not folded.

The Basement Rock Map of the United States (Bayley and Meuhlberger, 1968) shows the basement deepening toward the southeast side of the Valley and Ridge province, but this study of cross sections based upon the geometry of structures in outcrops, detailed geologic mapping, and expanded cross sections of Rodgers (1953a) suggest the opposite is true. Roeder (1977) also recently expanded the cross sections of Rodgers (1953a) with similar results. The many overlapping, thrust faulted units shown by Rodgers (1953b, Figure 3) are not necessary to satisfy the surface geology. With a northwestward tilt of the basement, gravity most likely played at least a part on deformation of miogeosynclinal rocks.

Some folds on the northwest side of the deformed miogeosyncline were produced by the upper plate of a thrust fault moving over a rise in a faulted surface. However, some major folds in the central portion of the Valley and Ridge Province, such as the Greendale syncline, originated prior to major faulting. This fold developed prior to the southeastern synclinal limb breaking and subsequently developing into the Saltville fault. In the area of detailed study, the Greendale syncline is truncated by the unfolded Saltville fault, thereby demonstrating that the syncline is older than the fault. Some of the large synclines, like the Greendale, possibly owe their incipient development to buckling of the crust from horizontally directed forces due to large convection heat cells in the mantle. Because of strata being much thicker in the axial region than on the northwest limb, Cooper (1964, p. 95) believed the Greendale syncline began development as a "great depositional downfold" in Ordovician time.

The synclines became asymetrical and break thrusts developed in their southeastern limbs because of horizontal compressional forces. Bailey Willis, as early as 1893, noted such structures in the Appalachians. Other folds and break thrusts restricted to the Paleozoic sedimentary rocks, i. e. those folds and faults lacking the basement deformations, may owe their regular frequency and great length to control exerted by the most competent thick unit in the stratigraphic section.

the Knox Group. Currie and others (1962), from their studies of horizontally compressed layered rock models, concluded that fold size and regular spacing between folds is controlled by the thickest competent layer. Watkins (1964) does not believe the regular spacing and great length of folds and faults of the Valley and Ridge Province can be attributed to basement influence because his studies of the gravity and magnetic features of the basement show little parallelism with Valley and Ridge Paleozoic structures.

Thrust faults most likely originally developed beneath the southeastern limbs of major anticlines in the incompetent strata beneath the Knox Group. As the orogeny progressed, the thrust faults merged into one, or a few, almost horizontal faults not far above the basement, a décollement. Willis (1893) noted that strata older than Rome Formation are not brought to the surface except on the east side of the deformed miogeosyncline. This fact plus the work of Buxtorf (in Cooper, 1961, p. 112) led to the development of the widely accepted concept of a single décollement in the Appalachian Valley and Ridge Province.

It is interesting to note that the Rome Formation in many areas is highly folded and faulted, especially in areas which are probably axial regions of folds. One can easily visualize the thick, competent Knox unit buckling with the incompetent, shaly Rome Formation filling the axial regions of the folds immediately above the massive competent basement below or above competent units, such as the Shady Dolostone. The cross sections (Figure 2) of major structures were modeled after features similar to those in outcrops.

Many folds and faults can be observed in outcrop within the area of detailed study. Fold-pair break thrusts have been observed in single outcrops. Three occur in road cuts of U. S. Highway 25-E near Collins Ridge. Folds are overturned to the northwest with an anticline on the southeast. The fault passes through a competent dolostone or sandstone bed of the Rome Formation and continued under the competent bed to the southeast and becomes parallel to bedding.

Folds are of the concentric type of flexural-slip origin, and incompetent units have filled low pressure areas in the axial regions by internal folding (some are chevron type) and faulting. Outcrops of such folds are abundant in abandoned railroad and road cuts of the shore of Cherokee Lake, 0.9 miles west of the intersection of Collins Ridge and U. S. Highway 25-E. They occur in the Nolichucky Shale and the rocks are thin beds of competent limestone and incompetent shale.

CONCLUSIONS

1. Northeast-trending sags probably developed in basement rocks during early Paleozoic time due to convection in the mantle. This likely resulted in a few local basins of deposition in the miogeosynclinal basin which later acted as the loci for synclines developed by horizontal

compression from the southeast.

2. Anticlines may have developed on the east side of these synclines. The Knox Group probably acted as the competent unit controlling size of the first anticlines and the position and size of many subsequent folds in the Valley and Ridge Province.

3. Based upon outcrop exposures of folds, major folds are of concentric geometry and developed by flexural-slip.

4. The larger synclines broke on the southeast limbs and resulting thrust faults merged into a few almost horizontal faults with a major décollement in the lowest incompetent unit above basement.

5. Geologic evidence does not require the basement to presently be deeper on the east side of the miogeosyncline, and this suggests a westward tilt of basement during major deformation allowing gravity to be the principal deforming force of the orogeny in the miogeosyncline.

6. Some folding on the west side of the miogeosyncline was caused by movement of rocks over an irregular fault surface.

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RATE OF MID-PALEOZOIC OROGENIC UPLIFT IN THE SOUTHERN APPALACHIANS

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ABSTRACT

Paleozoic orogenic uplift in the core of the southern Appalachians is widely acknowledged as having begun in Middle Ordovician time. Based on sedimentologic and stratigraphic information, the rate of uplift at the start of this uplift cycle during the Chazyan stage is calculated to have been about 8 cm per century. The first appearance of high-grade metamorphic minerals in the detrital suite of Valley and Ridge sedimentary rocks is in Mississippian time. Experimental data implies that the depth at which these high grade minerals develop is about 22 to 23 km; therefore, the average rate of uplift during the entire uplift cycle between Middle Ordovician and Mississippian time was about 2.4 cm per century. Data from other mountain belts pertaining to the pattern, duration, and rate of orogenic uplift are in harmony with those determined here. It is suggested that the rate of uplift may be approximated by an exponential decay equation. A graph of this form, adjusted to fit the calculated uplift rates, is presented. The "orogenic half-life" of this mid-Paleozoic episode is estimated to have been about 19 million years.

INTRODUCTION

Many writers in discussing tectonism and deformation in orogenic regions use generalized descriptive terms such as "episode of vigorous tectonism", "time of mild deformation", "culmination of uplift", or other similar phrases to convey an idea of the relative rate or intensity of earth movements. Graphic representations are also employed; these usually take the form of smooth to jagged curves that indicate time of relatively high or low rates or intensities of orogeny or its components such as metamorphism, deformation, uplift, and so forth. With the continuing refinements of geochronology, the time axis of these graphs may be carefully scaled and plotted, however, we are rarely able to assign numerical values to the other axis which indicates intensity or rate. Instead, the amplitudes of highs and lows on the curves are usually based on subjective judgments made by each worker according to his own personal concept of the relative values. Though these descriptive techniques are certainly useful, quantifying the concepts and providing some numerical values for rates and intensities of orogeny would be scientifically desirable. This article focuses on the North Carolina-east Tennessee section of the southern Appalachians from Middle Ordovician to Mississippian time. Later events, such as the Acadian or Alleghenian orogenic episodes, are not considered here. Factual data for the southern Appalachians during this time, along with supportive information from other areas, are presented which permit calculation of rates of uplift--the vertical component of orogeny.

In the geologic development of the Valley and Ridge, it has long been realized that a major tectonic and sedimentologic discontinuity exists at the Early Ordovician-Middle Ordovician boundary. The older pattern, which had persisted since at least late Precambrian, involved continuous deposition into a gradually subsiding epicontinental shelf area with the detrital components being derived from a cratonic source located westward of the depositional basin. Succeeding this phase there was an abrupt change in behavior; the eastern part of the original basin was uplifted, and a portion of the area between the uplifted region and the craton began rapid subsidence. As a result, the provenance and direction of transport of detritus became entirely different. No longer did the western craton supply the sedimentary debris; instead, streams and rivers flowed off land areas located to the east or southeast and transported their muddy, sandy loads westward. This new, eastern source region, once identified as "Appalachia", is nowadays generally given a more genetic term. Hatcher (1972, p. 2752) has applied the phrase "tectonic land" to this uplifted area and succinctly describes the geotectonic setting; "The uplift of a tectonic land over the zone of maximum mobilization and thermal activity provided a provenance area for syntectonic sediments." It is, of course, the deformed and metamorphosed roots of the "tectonic land" that constitute the present Blue Ridge and Inner Piedmont terrane. Nearly all studies, from the most detailed to the most casual, find that compression-produced structures, overturned to recumbent folds, thrust or reverse faults, and widespread shear features, are characteristic of this terrane. Thus the reasonable inference is drawn that the regional stress system responsible for the tectonic uplift was also compressional. King (1950, p. 659) expressed this conclusion by identifying the newly uplifted source area as "fold ridges raised in the interior zones of the mountain system." The overall uplift then, is clearly related to fundamental, mountain making forces of the Appalachian system; consequently, numerical values assignable to the uplift rate are at once a measure of this aspect of orogeny.

Acknowledgments

The author is thankful to several friends who kindly made available some of the heavy mineral information incorporated in Table I. These include Jerry Bundy, Ken Hasson, Larry Larson, and W. D. Lowry. I am also indebted to Carl Merschat, my colleague and careful coworker in the Blue Ridge, for continually helping clarify and sharpen my own ideas through our many geologic discussions and conversations.

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AVERAGE RATE OF UPLIFT

A direct approach for calculating uplift rates is to determine just how long it took for specific minerals whose burial depth is closely known to reach the earth's surface as a result of uplift. Some minerals suitable for this purpose commonly form as the result of high-grade metamorphism of pelitic material. Based on experimental data we may infer that the metamorphic assemblage of kyanite, sillimanite, muscovite, quartz, biotite, and feldspar, which is now exposed along the trace of the kyanite-sillimanite isograd, developed under a minimum pressure of about 6 kilobars. (See compilation plots by Dallmeyer, 1975, p. 453; Jansen and Schuling, 1976, p. 1248). Considering the average specific gravity of the overlying rocks to be about 2.7, the minimum depth of burial at the time of formation of this assemblage was probably about 22 1/2 kilometers.

The data presented in Table I reveals that the especially diagnostic aluminum silicate mineral, sillimanite, and kyanite, occur sparingly as detrital grains in the older Paleozoic rocks. These grains are presumed to have formed in pre-Paleozoic time and have been recycled and reworked into successively younger strata. In contrast, sillimanite and kyanite as well as staurolite and garnet become common, widespread detrital constituents following Devonian time. The interpretation here is that most of these grains were derived directly, or nearly so, from the just emerging high-grade, metamorphosed core of the southern Appalachians.

The conclusion then is that between the beginning of middle Ordovician time (estimated at 455 million years ago) when uplift commenced, and the beginning of Mississippian time (about 360 million years ago) (Lambert, 1971, p. 22) when kyanite and sillimanite-bearing rocks were first exposed, at least 22 1/2 kilometers of uplift had occurred. Thus, the average rate of uplift during this interval was about 2.4 cm per century.

Table I. Compilation of detrital heavy mineral analyses. Note widespread presence of metamorphic indicator minerals in Late Paleozoic strata.

Age	Stratigraphic Unit	Minerals Identified [(a)] Metamorphic Indicators	Other Minerals	Reference	Remarks
Pennsylvanian	Pottsville Group	kyanite	dumortierite (?), corundum, garnet, hornblende, monazite, R,I,T,Z	Larson, 1971, written comm.	dumortierite (?) may possibly be kyanite
			corundum, garnet, spinel, R,I,T,Z	Barnes, 1954	
			epidote, garnet, R,I,T,Z	Hill, 1951	
			corundum, garnet, R,I,T,Z	Shekarchi, 1951	
Mississippian	Parkwood Formation	staurolite	garnet, hornblende, monazite, orthopyroxene, R,T,Z, opaques	Whisonant, 1971	also slate, phyllite, and schist fragments
	Loyalhanna Limestone	kyanite, sillimanite, staurolite	garnet, R,T,Z	Adams, 1970	
	Grainger Formation	andalusite, kyanite, sillimanite, staurolite	augite, epidote, garnet, hornblende, monazite, R,T,Z	Hasson, 1976, written comm.	includes data from D. K. Sanders
		kyanite, sillimanite	garnet, hornblende, R,I,T,Z	Wiener, unpub. data	
			garnet, epidote, I,T,Z	Hill, 1951	
Devonian	Chattanooga Shale		garnet, R,T,Z	Bates and Strahl, 1957	
	Third Bradford sand (Chemung Group)	andalusite, kyanite, staurolite	aegirine-augite, diopside, epidote, hornblende, hypersthene, spinel, sphene, tremolite, zoisite, R,I,T,Z	Krynine, 1940	also slate and phyllite fragments
	Oriskany Sandstone		hornblende, R,I,T,Z	Bundy, 1976, pers. comm.	
	Ridgeley Sandstone		garnet, R,I,T,Z	Harris, 1972	
Silurian	Keefer Sandstone		R.T,Z	Folk, 1960	also slate to phyllite fragments
	Rose Hill Formation		R,T,Z	Folk, 1960	also chlorite-sericite slate to phyllite
	Tuscarora Formation		R,I,T,Z	Folk, 1960	also sericitic clay slate
			amphibole, garnet, R,I,T,Z	Yeakel, 1962	
		kyanite [(b)]	R,I,T,Z	Harris, 1972	
	Massanutten Sandstone		monazite, Z	Harris, 1972	
			garnet, R,I,T,Z	Cundiff, 1951	
	Clinch Sandstone		R,I,T,Z	Hill, 1951	
Late Ordovician	Juniata and Bald Eagle Formations		clinozoisite, sphene, I,T,Z	Yeakel, 1962	also low-grade metamorphic rock fragments
	Martinsburg Formation		amphibole, garnet, pyroxene, sphene, T,Z	McBride, 1962	also phyllite fragments
Middle Ordovician	Bays Formation	andalusite [(c)]	augite, corundum, epidote, garnet, gold, hornblende, spinel, topaz, zoisite, R,I,T,Z	Cummings, 1965	
			epidote, garnet, sphene, R,I,T,Z	Nelson, 1955	
	Sevier Shale		garnet, I,T,Z	Nelson, 1955	

Table I. Compilation of detrital heavy mineral analyses. - Continued.

Age	Stratigraphic Unit	Minerals Identified [a] Metamorphic Indicators	Other Minerals	Reference	Remarks
Middle Ordovician	Athens Shale		epidote, sphene, I,T,Z	Nelson, 1955	
	Holston Formation		garnet, hornblende	Ayrs, 1933	
	"Fincastle" [d] conglomerate		clinozoisite, I,T,Z	Wiener, unpub. data	W.D. Lowry reports kyanite and sillimanite not present, written comm., 1976
Early Ordovician	Uppermost Knox Group		R,I,T,Z	Hill, 1951	probably Mascot Fm.
			epidote, sphene, I,T,Z	Nelson, 1955	probably Mascot Fm.
	Kingsport Formation		garnet, sphene, R,I,T,Z	Nelson, 1955	
	Chepultepec Dolomite	sillimanite	cassiterite, dumortierite, epidote, garnet, hornblende, periclase, sphene, topaz, tremolite-actinolite, zoisite, R,I,T,Z	Cummings, 1959	
Early Cambrian	Rome Formation		garnet, R,I	Nelson, 1955	
	Chilhowee Group		I,T,Z	Swingle, 1949	
		kyanite [b]	R,I,T,Z	Harris, 1972	
			R,I,T,Z	Hill, 1951	
Precambrian	Walden Creek Group		epidote, sphene, I,T,Z	Hamilton, 1961	
	Great Smoky Group [e]		epidote, R,I,T,Z	Carroll, et al., 1957	
			epidote, sphene, I,Z	Hadley, 1970	
	Snowbird Group		R,I,T,Z	Carpenter et al., 1966	

(a) Apatite and magnetite, although commonly reported, are not included in this tabulation. Abbreviations: R = rutile; I = ilmenite, leucoxene, or anatase; T = tourmaline; Z = zircon

(b) Kyanite may have been a contaminant in sample analyzed (Harris, written comm., 1976).

(c) Of 2,600 grains identified from 13 different samples, one grain of andalusite is recorded.

(d) Informal nomenclature, see Kellberg and Grant (1956) for location.

(e) May be of Early Paleozoic age (Wiener, 1976).

RATE OF UPLIFT DURING MIDDLE ORDOVICIAN TIME

Data presented in the often-cited article by Kellberg and Grant (1956) provide key information for determining the uplift rate during a much briefer span of geologic time in this same region. Kellberg and Grant mapped and described occurrences of unique coarse conglomerate units found at a number of places from central Virginia, across East Tennessee, and into Georgia. These sandy to bouldery beds are found in rocks of Middle Ordovician Chazyan age, stratigraphically higher than the Lower Ordovician-middle Ordovician boundary, and well below early Trenton bentonite-bearing beds in East Tennessee dated at 447 $\pm$ 10 million years (Adams et al., 1960, p. 638). Pebbles and cobbles in the conglomerates were examined by Kellberg and Grant who report the presence of representatives from all the older, underlying sedimentary formations down to and including quartzite from the basal Cambrian Chilhowee Group. Since widespread sandstone beds at about the same stratigraphic level as the conglomerate layers are notably feldspathic (e. g. Tellico Formation; Neuman, 1955, p. 170), subaerial erosion must have continued downward well into strata containing abundant feldspar. The highest stratigraphic unit that meets this requirement is the lower portion of the Chilhowee Group (the Unicoi Formation of northeastern Tennessee and Virginia and its lateral equivalent further south, the Cochran Conglomerate). Therefore, based on known, well established thickness data (Swingle, 1966), the minimum amount of strata which must have been uplifted and eroded in order to account for the diversity of material found in the conglomerate and associated beds is about 3 1/2 kilometers (Knox Group - 1000 meters; Conasauga Group - 650 meters; Rome Formation - 500 meters; Shady Dolomite - 350 meters; upper part of the Chilhowee Group - 1000 meters). Uplift, of course, did not merely raise the Chilhowee beds to sea level; it must also have brought them up enough additional distance to have created a highland area. Although the exact elevation is unknown, perhaps a reasonable guess is 500 meters, thereby making the total uplift approximately 4 kilometers. The duration of this uplift was within the span of the Chazy; unfortunately our knowledge of the absolute duration of this interval is not very precise; however, it is likely in the order of only 5 million years or so. Using this value, the rate of uplift during Chazyan time was about 8 cm per century.

OTHER ESTIMATES OF UPLIFT RATE

Recently Dallmeyer (1975) has calculated the rate of uplift in the Bryson City area of the central Blue Ridge of North Carolina. He has made clever use of the concept that cooling associated with regional uplift will differentially set the potassium-argon radiometric "clocks" of coexisting biotite and hornblende. By estimating the depth at which the "clock-setting" occurs for each of the two minerals, he was able to calculate the amount of uplift that occurred during the interval between the setting of the hornblende clock and the setting of the biotite clock. Essentially Dallmeyer developed a value for the geothermal gradient at the time of high grade metamorphic conditions during the early Paleozoic and then used this value to calculate the depths at which biotite and hornblende cooled enough to attain closed system behavior with respect to argon. With this approach, he obtained an average uplift rate of about 1. 0 cm per century between 415 and 345 million years ago. I view this figure with considerable caution mainly because of two implicit, unproven assumptions about the geothermal gradient that Dallmeyer had to make. These are: (1) the geothermal gradient is presumed to have been constant from the surface downward to about 25 to 30 kilometers at the time of development of high-grade metamorphic minerals; (2) the geothermal gradient is also presumed to have remained at this constant value for the following 100 million years or so.

In other parts of the world, uplift rates between 2 and 8 cm per century are cited for the Alps, Chilean Andes and Scottish Caledonides (Dewey and Pankhurst, 1970, p. 380). This range of values, of course, fits very nicely with the uplift rates of 2.4 and 8 cm per century calculated herein for the southern Appalachians. These different mountain systems range considerably in both time of development and geographic location; the fact that their uplift rates fall within rather narrow limits appears to me to warrant the inference that some fundamental physical attribute of rocks, likely their rheodynamic properties, is a worldwide primary control of the orogenic process.

DURATION OF OROGENIC PULSES

Although not sufficient by themselves for calculating rates of uplift, estimates of the duration of orogenic pulses indicate that the time span for the notably rapid uplift phase and associated intense metamorphic or deformational conditions is relatively brief. Earlier in this article it was estimated that much of the rapid uplift of the Blue Ridge and Inner Piedmont during the Middle Ordovician was accomplished in about 5 million years. This episode is intimately related to a wedge or delta-like accumulation of clastic sediments whose thickest exposed occurrence is centered in east Tennessee in Blount, Monroe and Sevier Counties. Rodgers (1953, p. 94) incidentally, has termed these coeval orogenic effects "the Blountian phase of the Taconic orogeny". Younger but quite similar clastic wedges are well known at other places in the Appalachians. Information for three of these complexes; the Late Ordovician-Silurian Queenston delta, the middle-late Devonian Catskill delta, and the Late Mississippian-Pennsylvanian Mauch Chunk-Pottsville wedge, is analyzed by Meckel (1970, esp. p. 66). Based on strictly sedimentological evidence, he deduces a common pattern for movement of the source area of each of these three wedges. The pattern involves a period of very active uplift and erosion followed by relative quiescence. Using Meckel's model for these mid-Paleozoic events and then setting them into the chronologic frame of the absolute time scale, we see that the time span for each of the episodes of source area uplift is not more than some 20 to 25 million years, with the most active uplift actually taking place during about half of this span.

Naylor (1971), working with metamorphosed rocks in eastern Vermont, has also come to the conclusion that the more intense episides of orogeny take place in a relatively brief interval. His rubidium-strontium isotopic data from granitic bodies that cross-cut previously metamorphosed, datable metasedimentary strata permit a maximum of not more than 30 million years for deposition, burial to depths exceeding

12 kilometers, and development of the dominant deformational structures and regional metamorphic effects. Naylor (1971, p. 559) comments that the actual time span for the orogenic phase must be considerably less.

Dewey and Pankhurst (1970, p. 383), after a careful review of radiometric and stratigraphic information for the Scottish Caledonides, conclude "... that the entire sequence of deformation and metamorphism occurred in a relatively short climactic episode, between 510 and 480 m. y. " and they go on to say "steady-state isostatic and thermal conditions were not finally attained until Middle Devonian time... 100 m. y. after deformation and metamorphism" (p. 384). Thus Dewey and Pankhurst's conclusions with respect to the lengths of time involved are similar to those derived here for the southern Appalachians.

GRAPHICAL REPRESENTATION

Although orogenic uplift rates for two different time periods have been calculated and some estimate for the duration of orogenic phases has been presented, this information in itself if not sufficient to empirically define a rate-of-uplift vs. time curve. To do so will require much more information; unfortunately numerical data of this sort are not readily derived. In lieu of additional facts, perhaps one may be justified in speculating on this question. To me, the orogenic pattern described in the preceding paragraphs suggests a relatively steady-state condition disturbed by a damped perturbation. If so, the relation between time and rate of change, or uplift, following the perturbing event may be approximated by an exponential decay curve or rate law equation of the well known form:

$$\text{Rate} = a\,e^{-(\lambda t)} + b$$

Rate = instantaneous uplift rate after some interval of time

t = time elapsed since start

e = base of natural logarithms

λ = half life

a, b = constants

A curve satisfying this general form as well as accommodating the uplift values for the southern Appalachians calculated previously is presented in Figure 1. (The author determined this curve by trial and error graphic integration; more elegant arithmetical techniques also yield the solution). The concept of half-life is inherent in equations and graphs of this form and inspection of Figure 1 reveals that the uplift rate is halved about every 19 million years. In other words, we might think of this episode as having an "orogenic half-life" of 19 million years. By summing the area beneath the rate curve, one may find values to construct a curve showing the total uplift at any time. Figure 2 presents such a curve. Both graphs clearly indicate that the mid-Paleozoic orogenic pulse, the subject of this paper, had essentially run its course by Late Devonian or Mississippian time.

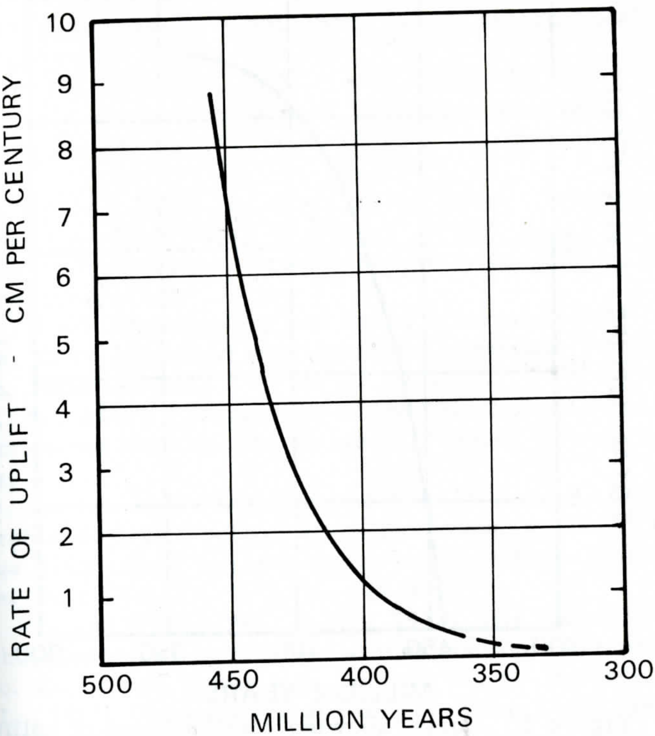

Figure 1. Rate of uplift of source area during mid-Paleozoic.

CONCLUSION

The information marshalled in this report leads to the conclusion that Paleozoic orogenic uplift between Middle Ordovician and Mississippian time in the southern Appalachians has followed a characteristic pattern; that of a short-term orogenic pulse or phase of rapid uplift which decays into a much slower, longer-term upward movement. For the Blue Ridge-Inner Piedmont terrane the uplift during the most rapid phase, in Chasy time, was about 8 cm per century. The average rate during the longer interval from Middle Ordovician to Mississippian was about 2.4 cm per century. These values are in harmony with rates determined by other geologists for some of the other mountain systems of the world. Based on the hypothesis that the rate of uplift may be approximated by an exponential decay equation or curve, an "orogenic half-life" of about 19 million years is deduced.

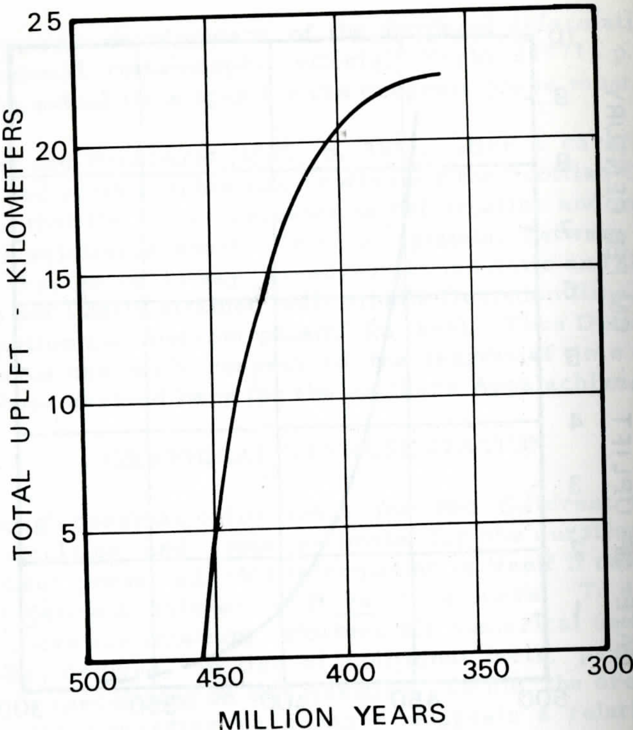

Figure 2. Curve showing total amount of uplift of source area.

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